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DEPARTMENT OF ENERGY, MINES AND RESOURCES

ECONOMIC GEOLOGY

REPORT No. 24

GROUNDWATER IN CANADA

Officers of the Geological Survey of Canada

Edited by I. C. Brown

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1967

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Officers of the Geological Survey of Canada
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DEPARTMENT OF
ENERGY, MINES AND RESOURCES
CANADA



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PREFACE

For many years the Geological Survey has been concerned with groundwater, and as far back as 1875 was requested to conduct an exploratory drilling program to obtain water for railway construction on the Prairies. After these early beginnings there was a long period in which the main emphasis was on data-gathering and well-recording. This phase continued up to the present decade, which has seen the development of many new instruments and techniques and the application of the resources of geological, physical, chemical, and biological sciences to the complex problems of hydrology and groundwater flow.

Water has come to be rightly regarded as a major factor in the economy of the nation. This report presents a summary of data and provides an assessment of the present state of knowledge of groundwater studies in Canada.

Y. O. FORTIER,
Director, Geological Survey of Canada

OTTAWA, January 24, 1966

ECON. GEOL. REPORT Nr. 24 — DAS GRUNDWASSER IN KANADA. Zusammengestellt in der Geological Survey of Canada.
Von I. C. Brown

Gibt einen Überblick über die Grundwasservorkommen und beschreibt die Hydrogeologie der bedeutendsten hydrogeologischen Gebiete Kanadas.

ЭКОНОМИЧЕСКАЯ ГЕОЛОГИЯ, ДОКЛАД № 24 Служащие Геологического управления Канады, ред. И. Ч. Браун. ГРУНТОВЫЕ ВОДЫ В КАНАДЕ

В этом докладе дается сводка о распространении грунтовых вод и описывается гидрогеология главных гидрологических областей Канады.

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GROUNDWATER IN CANADA

Abstract

This report is the first detailed account of the occurrences, flow systems, and chemistry of groundwater in Canada. It relates the groundwater to the geological materials through which it flows and to the hydrologic cycle, and discusses comprehensively the principles and methods of hydrogeology and the methods, techniques, and formulae required for aquifer exploration and development.

Canada contains six major hydrogeological regions. Much groundwater data are available for the Interior Plains and St. Lawrence Lowlands hydrogeological regions, where groundwater is used extensively. Less data are available for the Appalachian, Canadian Shield, and Cordilleran hydrogeological regions, where surface waters are adequate to provide present demands. Little data are available for the northern region, where the population is sparsely distributed and permafrost influences the role of the groundwater.

Résumé

Le présent rapport est le premier exposé détaillé des nappes, des systèmes d'écoulement, et de la composition chimique des eaux souterraines au Canada. Il relie les eaux souterraines aux matériaux géologiques à travers desquels elles circulent ainsi qu'au cycle hydrologique. Il traite en détail des principes et des méthodes d'hydrogéologie, ainsi que des méthodes, des techniques et des formules utilisées dans la recherche et l'exploitation des nappes aquifères.

Le Canada renferme six grandes régions hydrogéologiques. Il y a abondance de données sur les eaux souterraines des plaines Intérieures et des Basses-Terres du Saint-Laurent, où se fait un usage considérable de ces eaux. Pour les régions hydrogéologiques des Appalaches, du Bouclier canadien et des Cordillères, où les eaux superficielles se trouvent en assez grande quantité pour répondre à la demande actuelle, les données disponibles sont moins nombreuses. Quant à la région septentrionale, où la population est clairsemée et où le pergélisol influe sur le rôle des eaux souterraines, les données sont plutôt rares.

Chapter I

INTRODUCTION

by I. C. Brown

In Canada, groundwater contributes about 10 per cent of the water supplied by municipal water supply systems serving communities with a population of more than 1,000, and an equal or greater amount of that used in individual homes. (*See* Table I.) In addition, large, though unrecorded, quantities of industrial and irrigation water are obtained from groundwater. In order to be advantageously and economically utilized such a vital renewable resource needs to be fully understood. If one judges from enquiries received by the Geological Survey at all levels from homeowners to groundwater consulting engineers, few people in Canada understand groundwater well and even hydrogeologists, who make an intensive study of it, admit that they are just beginning to understand its complexities.

At this stage in our study, a review of the status of our knowledge of the behaviour and occurrence of groundwater, particularly as related to the various hydrogeological regions of Canada, seems appropriate. The following is an attempt to provide a summary in general terms of our present knowledge of the behaviour of groundwater along with a more detailed presentation of principles and formulas in Chapter II; to summarize our knowledge of groundwater in the various hydrogeological regions of Canada; and finally to discuss the problems involved in the exploration and development of groundwater resources.

History of Development of Groundwater Studies in Canada

The first investigation specifically mentioning groundwater was made by the Geological Survey of Canada in 1875 when exploratory drilling was done in the Northwest Territories, now Saskatchewan, to obtain water for railway construction. In a short report on drilling for groundwater at Deloraine and Morden, Manitoba, A. R. C. Selwyn in 1890 gave a summary of the groundwater possibilities from the Red River to the Pembina Escarpment. The first Geological Survey report fully on groundwater was in 1901 by F. D. Adams and O. E. Le Roy, on the artesian wells of Montreal Island. Mapping of the Milk River sandstone in southern Alberta in 1915 indicated that it should be an extensive artesian aquifer and the following year a drilling program substantiated this. In the 1920's and 1930's, reconnaissance mapping on the Prairies for oil, gas, and water produced some wide ranging but excellent reports. In 1935 a massive groundwater inventory of the wells in southern Saskatchewan and east-central Alberta was undertaken by the Geological Survey and two

Table I

Canadian Water Supplies

Province	No. of systems	No. supplying groundwater entirely or partly	Water supplied in 1,000's of gal. per day		Population served in 1,000's	Per capita supply gal. per day
			Groundwater	Total water		
Alberta	64	26	3,755	90,537	810	112
British Columbia	83	20	9,664	160,711	1,278	126
Manitoba	42	10	1,333	85,819	1,048	82
New Brunswick	23	16	10,185	57,019	250	228
Newfoundland	21	1	300	23,217	155	150
Nova Scotia	34	14	4,660	38,408	341	113
Ontario	294	120	97,824	669,638	6,443	104
Prince Edward Island	3	3	2,908	2,908	29	100
Quebec	272	81	28,156	573,660	4,286	134
Saskatchewan	47	33	10,691	51,348	254	202
Totals	883	324	169,476	1,753,265	14,894	Average 135
		Groundwater supplied % of total 9.7				

Water supplied by Canadian municipal water supply systems serving populations of 1,000 or more. Does not include large industrial or private supplies not depending on municipal utilities. (Compiled from statistics in Canadian Municipal Utilities, Waterworks Manual and Directory for 1963.)

Type of supply	Number of dwellings	Average occupancy	Estimated water supply gpd per dwelling	Estimated water supply 1,000's of gal. per day
Piped into dwelling	759,349	3	200	151,870
Not piped into dwelling	377,401	3	100	37,740
Totals	1,136,750			189,610

Estimated water supply of dwellings not supplied by municipal water supply systems. Sources of water unknown but probably largely from wells. (Compiled from statistics in Canadian Housing Statistics, 1962, Central Mortgage and Housing Corporation, March, 1963.)

hundred and seventy-four water supply papers were published from the results of this work. This type of well inventory work was continued on a smaller scale in Ontario, Quebec, Prince Edward Island, and on the Prairies, with increasing emphasis on the quantitative and chemical aspects and on analysis of older data until about 1959. Since then, due to the growth and cooperation of provincial groundwater agencies, this inventory type of study is no longer necessary, and the Geological Survey has concentrated its studies on groundwater flow systems and on research into the fundamental behaviour of groundwater. Well inventory data are now handled almost entirely by provincial agencies, many of whom require drillers to submit data on all wells drilled. Such data are kept readily available and are a vital source of information for anyone conducting groundwater research or looking for a groundwater supply (Table II).

Table II

Sources of Groundwater Information in Canada

British Columbia —Groundwater Division,
Water Investigations Branch,
Dept. of Lands, Forests and Water Resources,
Parliament Buildings,
Victoria, B.C.

Alberta —Groundwater Division,
Research Council of Alberta,
10302—107th Street,
Edmonton, Alta.

Saskatchewan —Water Rights Branch,
Saskatchewan Water Resources Commission,
S.P.C. Building,
Regina, Sask.

Manitoba —Water Control and Conservation Branch,
Dept. of Highways,
Norquay Building,
Winnipeg 1, Man.

Ontario —Groundwater Branch,
Ontario Water Resources Commission,
801 Bay Street,
Toronto 5, Ont.

Quebec —Groundwater, Gas and Petroleum Branch,
Dept. of Natural Resources,
1620 Boulevard de l'Entente,
Quebec City, Que.

New Brunswick —Mines Branch,
Dept. of Natural Resources,
P.O. Box 758,
Fredericton, N.B.

Nova Scotia —Nova Scotia Department of Mines,
Provincial Building,
Halifax, N.S.

Prince Edward Island—Dept. of Industry and Natural Resources,
Charlottetown, P.E.I.

Newfoundland —Mineral Resources Branch,
Dept. of Mines, Agriculture and Resources,
St. John's, Nfld.

General—Ottawa —Inland Waters Branch,
Dept. of Energy, Mines and Resources,
No. 8 Temporary Building,
Carling Avenue,
Ottawa, Ont.

Calgary —Inland Waters Branch,
Dept. of Energy, Mines and Resources,
3303—33rd Street NW,
Calgary, Alta.

Vancouver —Inland Waters Branch,
c/o Geological Survey of Canada,
Room 102,
326 Howe Street,
Vancouver 1, B.C.

Responsibility for administration and control of groundwater rests with the provinces, and most of them have established agencies to carry out these responsibilities and to assist in the development and understanding of their groundwater resources. The growth of the work done by these agencies has been parallel, to some extent, with that described for the Geological Survey of Canada, starting with data gathering and assistance in developing water supplies, and finally expanding to research in order to provide effective service and intelligent control.

The first provincial agency to enter the field was the Research Council of Alberta, which started field studies in 1930 and continued these at intervals until the Groundwater Division was formed in 1955. Since then, it has been carrying out continuous data gathering, assisting in locating groundwater supplies, doing groundwater mapping, and conducting a major research program into many aspects of groundwater.

Groundwater mapping was started by the Ontario Department of Mines in 1945 and continued until the Ontario Water Resources Commission took over the duties in 1957. Since then files have been kept on all wells drilled for water, assistance has been given to municipalities in finding and developing groundwater supplies, and research into aquifer characteristics and artificial recharge has been conducted.

The British Columbia Department of Mines started mapping for groundwater purposes in 1950 and continued until 1959 when the well inventory work was taken over by the Water Rights Branch, Department of Lands and Forests, now the Water Investigations Branch. They are making an inventory of all water wells in the province and conducting test drilling and groundwater mapping and research programs.

The Quebec Department of Mines, Geology Branch, started groundwater investigations in 1956, and in 1959 the Groundwater, Gas and Petroleum Branch of the Department of Natural Resources was formed to increase assistance in finding and developing groundwater supplies and to conduct research into aquifer characteristics. More recently the Branch started an active research program.

In 1959 the Geology Division of the Saskatchewan Research Council set up a groundwater research program that has been very active in mapping and research. Well data are gathered by the Water Rights Division of the Conservation and Development Branch, Department of Agriculture.

The Nova Scotia Department of Mines started collecting well data on a voluntary basis in 1960, and in 1963 started regular groundwater investigations. In 1964 drilling and research programs were also begun.

The Water Control and Conservation Branch of the Department of Agriculture and Conservation in Manitoba started collecting water data in 1961, and now carries on an active drilling campaign to assist communities in finding groundwater supplies, as well as conducting groundwater mapping and research programs.

Definitions of Commonly Used Terms

Groundwater as used here includes all water in the zone of saturation beneath the earth's surface except that water chemically combined in minerals.

The upper limit of completely saturated material is the *water-table* and for practical purposes it may be determined by the level of water standing in open holes

penetrating just into the saturated zone. It may more accurately be defined as the surface at which water is at atmospheric pressure (Figs. 1, 2). The ground may be wet to varying degrees due to the capillary rise of water and above this level it may be dry, but more commonly there is soil moisture present in amounts varying from very little to complete saturation.

The water-table may be at surface as along the shores of bodies of water or during wet periods when the ground is fully saturated, or, it may be at depths of many hundreds of feet as in mountainous arid areas. The lower limit of groundwater is determined by the depth at which pressure closes all openings in the rocks. This depth, which depends mainly on the structural strength of the rocks, has not yet been determined as the deepest oil wells still show signs of water as deep as 25,000 feet. The lower limit may also be determined by the critical pressure above which water ceases to be a liquid regardless of pressure. In any event the lower limit is of more academic than practical interest.

In a discussion of groundwater certain terms are commonly used, which are defined briefly below. Those wanting more comprehensive and quantitative definitions and references will find them in Chapter II.

Total Head. The measure of the hydraulic potential in water at a given point, or the height to which water will rise in a tube open only at the bottom. It is largely hydrostatic pressure but also includes pressure of motion. This is usually expressed as the height in feet above or below sea-level but may also be referred to any convenient datum level. Before one compares heads it is important to be sure that they are referred to the same datum.

Piezometric surface. A pressure surface or an imaginary surface passing through all points to which water would rise from a given depth or material. It is usually represented by contours as is topography on a topographic map. A piezometric map shows the pressure distribution at a given level or within some material which must be clearly defined for the map to be of any value.

Porosity. The ratio of water-filled spaces in a saturated rock or soil to the total volume of rock or soil. It is usually expressed as a percentage. It is only indirectly related to the size of the spaces and varies from almost zero in unfractured crystalline rocks to 20–40 per cent in sands and gravels or as high as 60 per cent in some clays and soils.

Permeability. Ability of a material to allow a fluid, in this instance water, to pass through it. It is controlled by the size and interconnection of the spaces in a rock or soil and by the viscosity of the water. In groundwater studies, it is normally expressed as the number of gallons of water per day that will pass through a cross-section of one square foot under a hydraulic gradient of one foot to the foot. A hydraulic gradient represents the difference in head in water between any two points, in this case a difference in head of one foot between two points one foot apart. Permeabilities vary greatly and may be from 1/100 gallon to 10 gpd per square foot in very fine materials, to values as high as 1,000,000 gpd per square foot in such permeable materials as clean gravels.

Transmissibility. The measure of the amount of water in gallons per day that will flow through a vertical strip one foot wide extending from top to bottom of the material being investigated and under a one foot to one foot hydraulic gradient.

Aquifer. Any water-saturated body of geological material from which enough water can be drawn at a reasonable cost for the purpose required. An aquifer is only a relative term determined largely by economics and is best illustrated by extreme examples. An aquifer in an arid prairie area required to supply water for a single farm may be adequate if it can supply 100 gpd. This would not be considered an aquifer by any industry looking for cooling water in volumes of 1,000,000 gpd. A common usage of the term aquifer is to indicate the water-bearing material in any area from which water is most easily extracted.

Artesian. This term is always used in conjunction with other terms to describe a particular condition, i.e., artesian aquifer, artesian basin, artesian well. Artesian conditions are present in any formation or series of formations in which the hydraulic heads cause an upward flow of water (Figs. 3, 4).

Aquiclude. A material of such low permeability that it effectively prevents any movement of water through it or reduces water movement to such a low value relative to the surrounding materials that it can be considered a barrier.

Recharge. Water added to the saturated zone from any source. This term is commonly combined with other terms to indicate some specific mode of recharge such as, recharge well, recharge area, artificial recharge.

Discharge. Any water removed from the saturated zone by any process. It is commonly used combined with other words as is recharge.

Storage. The amount of the water stored in a formation or specified part of a formation. Unless further defined by an associated term it refers to all the water in a formation and not only the part that can be recovered.

Hydrological Cycle

The hydrological cycle is the complete cycle of processes through which all water on earth is continuously moving (Fig. 1). The energy for this movement is provided by radiation from the sun, which continuously evaporates water from all land, water, vegetative, and animal surfaces into the atmosphere. The sun's energy plus the rotation of the earth continuously moves this atmospheric water to locations where conditions are such that it condenses in the form of either solid or liquid precipitation, and is pulled back to the surface of the earth by gravity. Some immediately evaporates again; some runs off over the surface into streams and lakes and eventually into the ocean; some sinks into the soil and is either extracted and transpired by plants or evaporates directly; some sinks more deeply into the ground and once it reaches the water-table becomes groundwater. This water, acting largely under the influence of gravity and the pressure of surrounding water, continually seeks areas of lower pressure or lower potential energy and follows very complex pathways through the

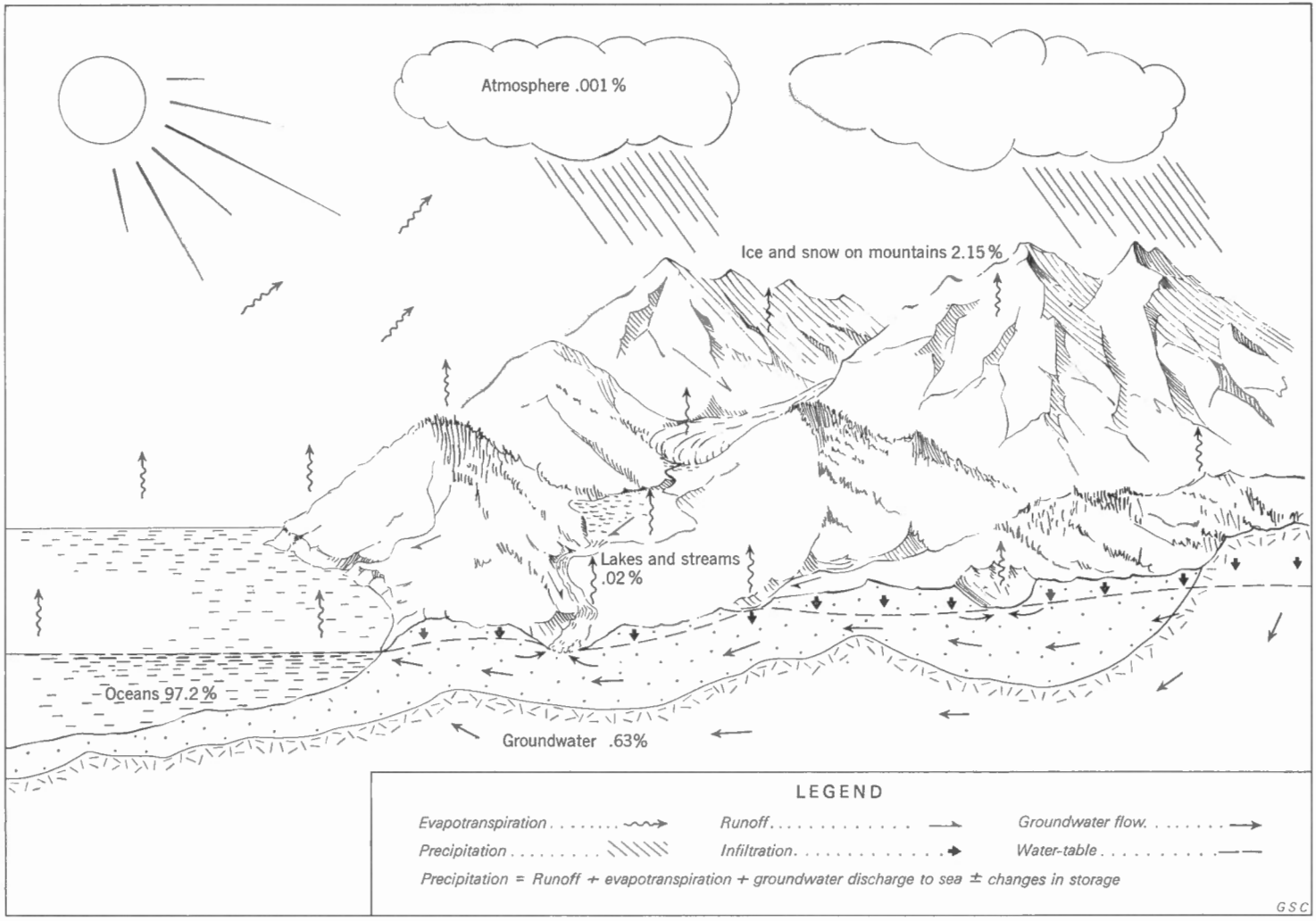


FIGURE 1. Hydrological cycle.

materials of the earth depending on the hydraulic gradient and the relative permeabilities of the materials. Some does not travel far before it returns to surface and joins the other parts of the cycle; some travels to great depths and remains underground for long periods; but all is eventually discharged onto the surface or into streams, lakes, or the ocean from where it is again evaporated and continues on around the cycle.

The hydrological cycle is always in balance and all parts are always in motion, even though the rate of movement may be so low that it may be considered zero for most practical purposes. The continuity of the hydrological cycle is best illustrated if we consider it in geological time terms. Ever since the earth's atmosphere became cool enough for water to condense, rain has been falling on the earth's surface. The first precipitation fell on barren rock and helped erode it so that deposits of sand, gravel, and silt were developed which could hold much more of the precipitation. These conditions continued for many thousands of millions of years before the advent of plant life some 350 to 400 million years ago. With the development of plant life came the development of soil, and the problem of soil moisture became much more complex; consequently, the recharge to groundwater became more complex and in total probably somewhat diminished.

Precipitation in Canada varies from more than 100 inches per year to less than 10 and at present the overall average is about 20 inches per year. Geological history shows that precipitation must have been considerably greater at times and considerably less at others; thus, the amount that eventually became groundwater must also have varied greatly. If we assume conservatively that only one half inch per year becomes groundwater, recharge then in a mere 1,000,000 years would add up to almost eight miles of solid water. Obviously all this water cannot just sink into the ground and disappear, but has to keep moving through it and must be discharged as fast as it is recharged over geological time periods, otherwise the ground surface would be saturated continually.

By far the largest volume of sedimentary rocks was deposited in the ocean even though these rocks now occur raised far above sea-level. Analyses of water from these rocks and particularly of deeper water as encountered in deep oil wells show salt values sometimes as high or even higher than that of sea water. However, the types and ratios of different salts and isotopes indicate that much, if not ¡all, of the original sea water has been moved out of them (Clayton, 1964). This ind cates movement of groundwater to very great depths.

Groundwater Fluctuations

If we consider precipitation as the ultimate source of all groundwater, then it is obvious that the source is very irregular both in time and in location, and the scale of this irregularity can vary from continent-wide changes to very local showers or even a single raindrop. Variations in time can be extremely long and some are measurable only in geological time. For example, there is evidence that many desert areas were once covered with vegetation and supported considerable animal life, and, conversely, rocks underlying some areas today indicate that they were laid down under desert conditions even though there is a plentiful vegetative cover and considerable rainfall now.

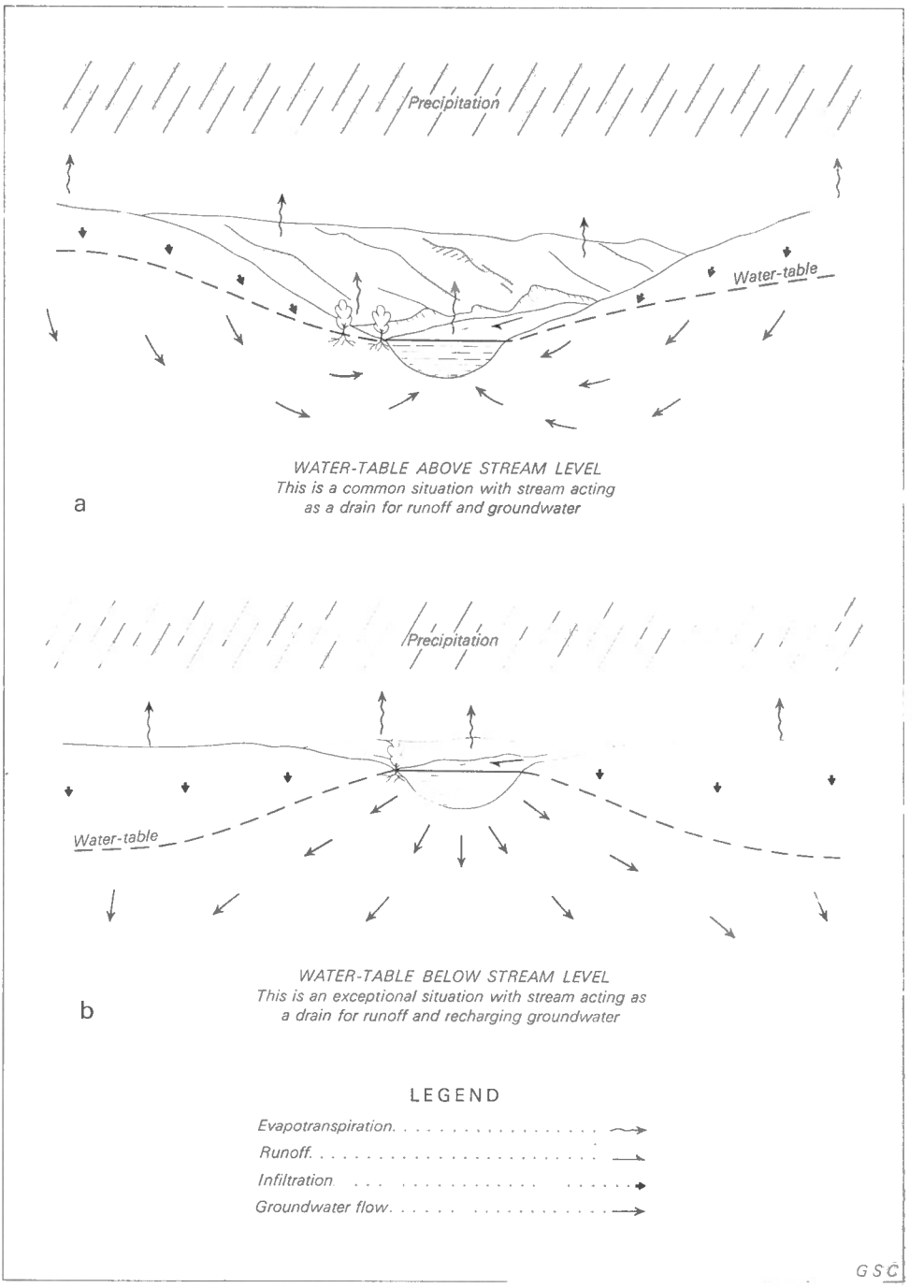


FIGURE 2. Hydrological cycle near: a, influent stream; b, effluent stream.

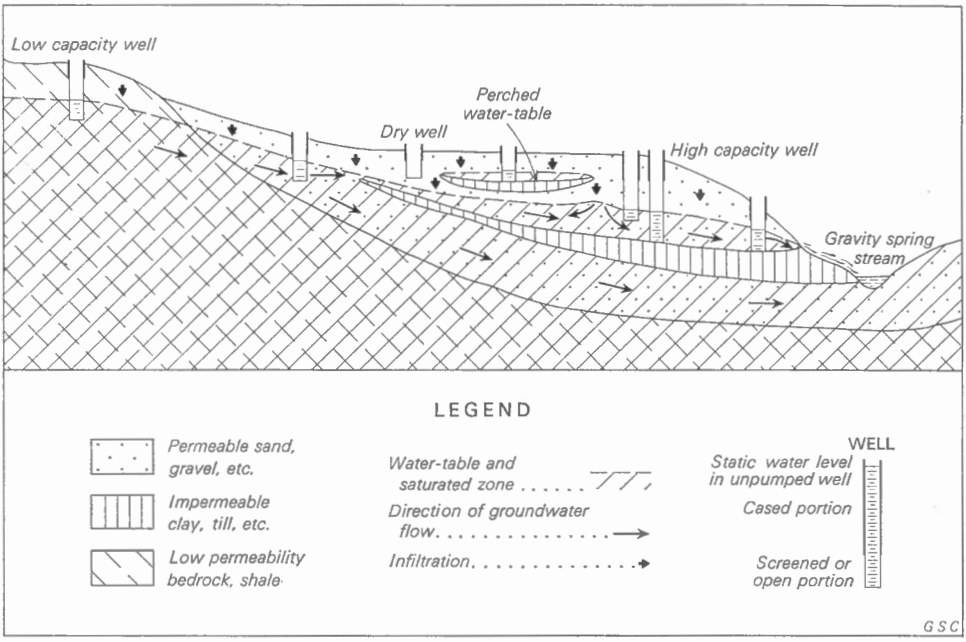


FIGURE 3. Water-table conditions (unconfined aquifers).

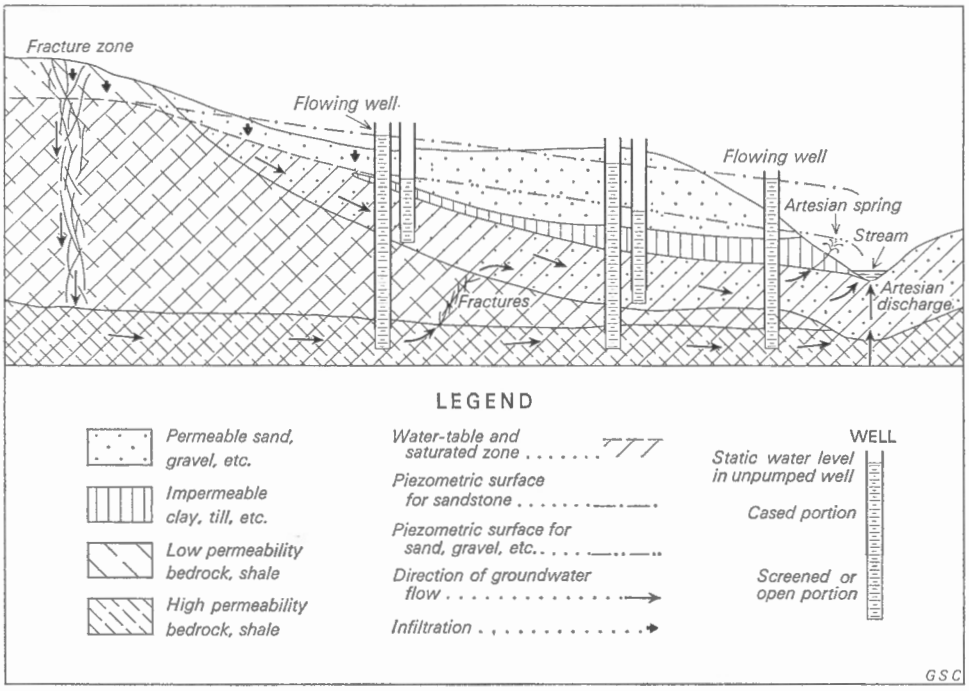


FIGURE 4. Artesian conditions (confined aquifers).

Rates of movement and fluctuations in surface water are slower and less erratic than those of atmospheric water because the storage of water in streams and lakes tends to smooth out the variations. Variations in groundwater flow in both time and space are considerably slower and more uniform than those of either surface or atmospheric water. The rates of groundwater flow depend on the hydraulic gradient and the permeability of the materials, and are commonly very slow as the hydraulic gradient is usually considerably less than the surface slope in areas of high relief and is, of course, very low in flat areas. Actual rates of movement vary from zero to a maximum of more than 100 feet per day but a normal range is more in the order of five feet per year to five feet per day (Todd, 1959). These low velocities combined with the vast storage available in the geological materials of the earth combine to reduce the variations in groundwater flow to a minimum. Thus, while the water-table may fluctuate comparatively rapidly, depending on the precipitation, any groundwater flow system of a reasonable length will vary much more slowly. The vast amount of water stored in the ground is about 30 times that stored in streams and lakes, but is still only a little less than a quarter of that stored in ice and about one hundred and fiftieth of that in the oceans (Fig. 1).

Variations from place to place in the average depth of the water-table below surface are much greater than the changes due to its continual fluctuations. Groundwater is present probably everywhere below the earth's surface but in some areas may be so deep that it has not yet been found or the materials may be of such low permeability and porosity, that it is almost impossible to detect. To both the user who wishes to extract water and to the hydrogeologist who wishes to study the hydrological properties of the water-saturated materials, the most important fluctuations are the irregular but recurring ones due to natural and artificial changes in conditions. These fluctuations vary greatly in magnitude and timing and are due to many causes (Figs. 5, 6).

Small rapid variations in the water level of wells in confined aquifers can be caused by any force that affects the materials in which water occurs, such as vibrations from traffic, earthquake forces, crustal tide forces, air pressure changes. These short term fluctuations are usually small and are measured in inches or fractions of inches, though those due to earthquakes have measured as much as a few feet. Because water is almost incompressible, these pressure changes should be transmitted long distances almost instantaneously, but due to the elasticity of the geological materials in which the water lies most of them are soon dissipated. As earthquake shocks are of great magnitude they may show up in fluctuations in the water-table over many thousands of miles or even right around the world. Longer term fluctuations are due to variations in supply of recharge water and are more likely to be seasonal. They vary in magnitude depending on the rate of movement in the flow system and the frequency and amount of recharge; in most areas of Canada this variation ranges from 5 to 20 feet. Natural pressure changes may not show up as fluctuations in deep wells until some considerable time after the change took place, because of the elasticity of the geological materials and the vast volumes involved.

Fluctuations in the water-table are also caused by man's interference with any part of the flow system by pumping, irrigation and drainage, altering river or lake

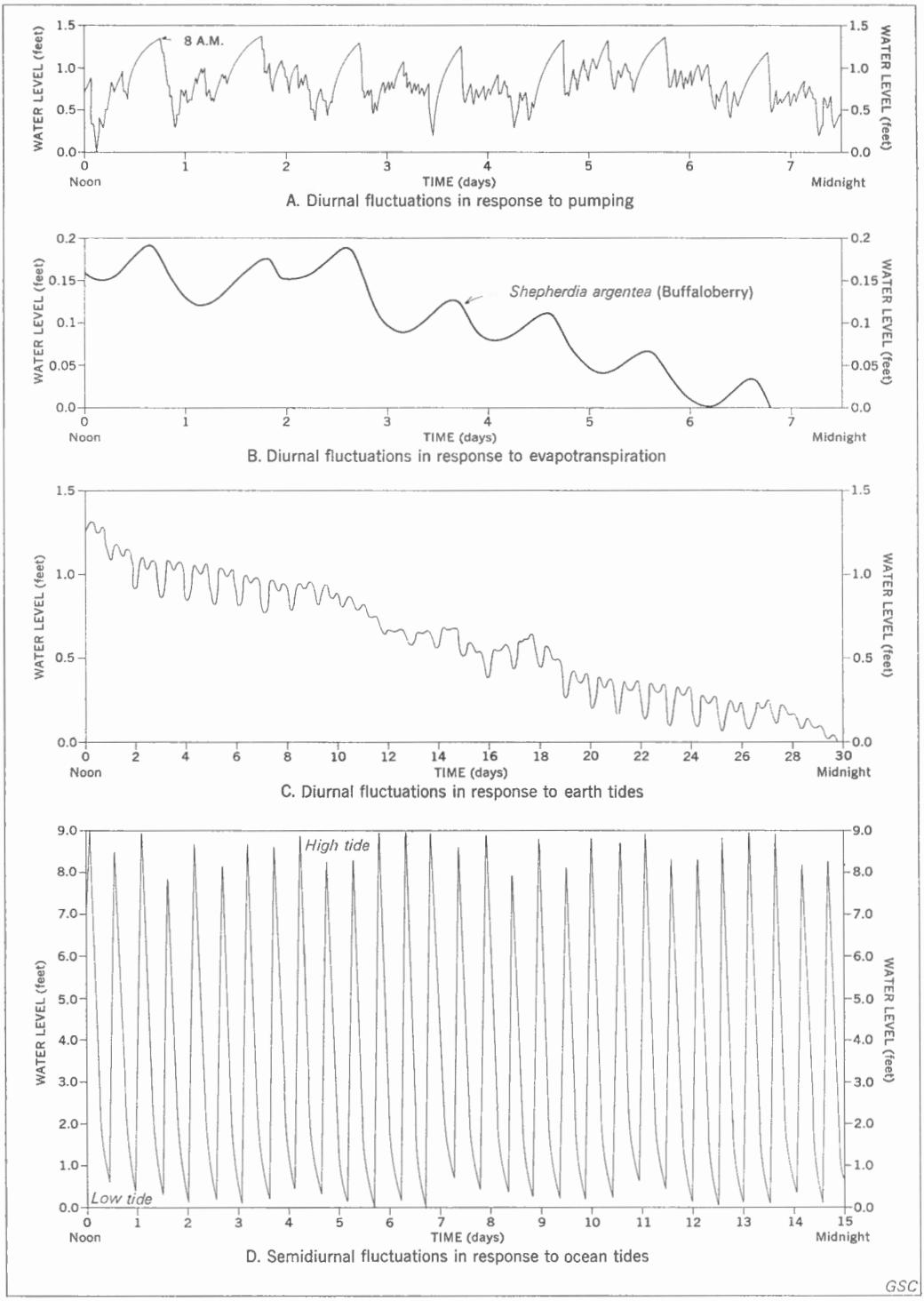


FIGURE 5. Diurnal and semi-diurnal fluctuations in groundwater levels.

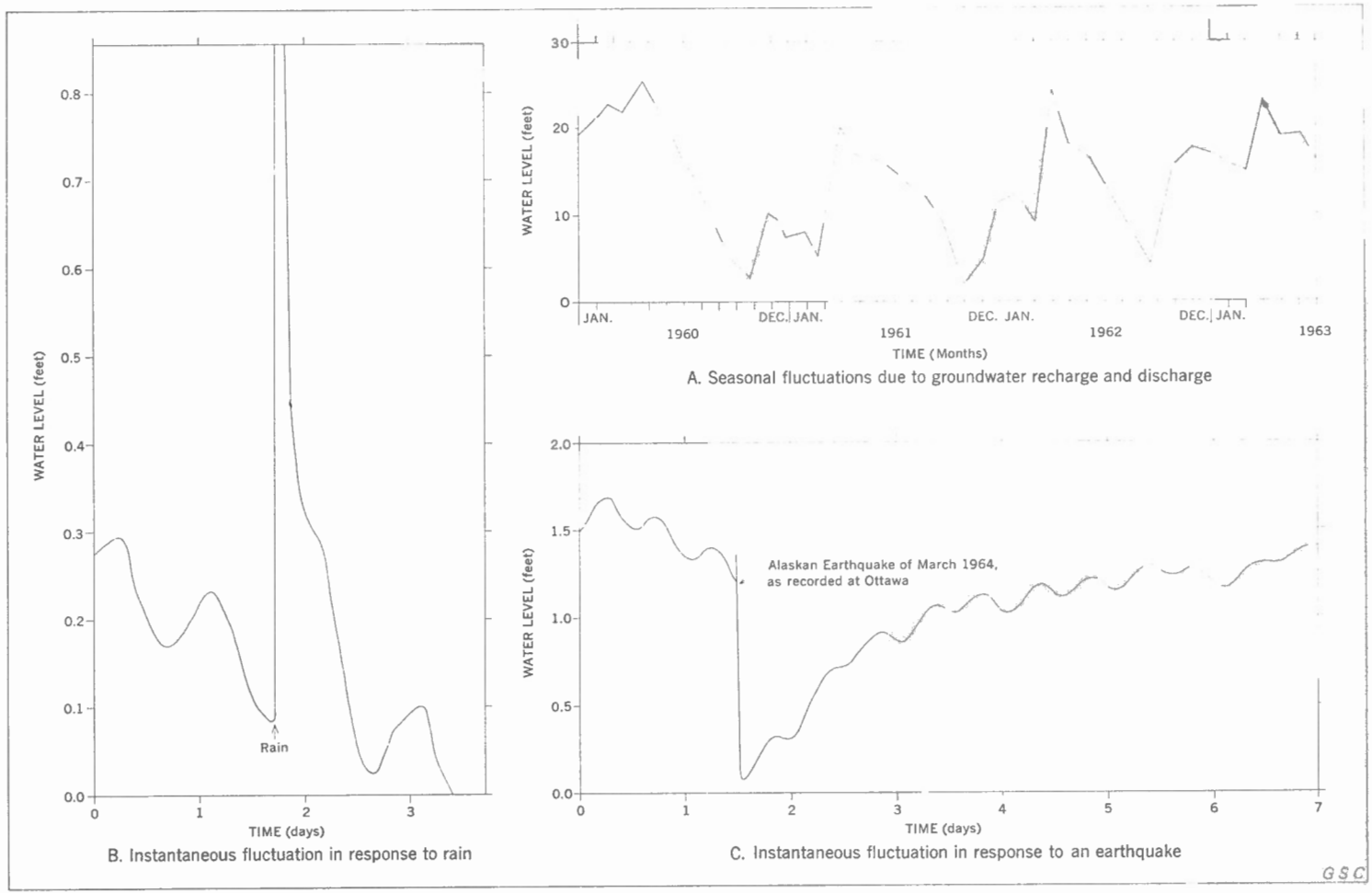


FIGURE 6. Seasonal and instantaneous fluctuations in groundwater levels.

levels, changing or removing the natural cover, and many other causes. In Canada, such fluctuations are known to vary in amplitude from a few inches to many tens of feet and in extent from a few hundred feet to many miles. In other areas of the world where man's interference has been more extensive they are correspondingly greater.

In general throughout Canada, recharge is greatest and groundwater levels are highest during spring break-up (Fig. 6). From this peak, they fall gradually to a low late in the summer; only very extensive rainy periods during the summer will make much difference as most of summer precipitation is picked up by vegetation and returned to the atmosphere. The lowest levels are reached in the early autumn, but the autumn rains provide considerable recharge as much of the plant life has stopped growing and very little water is being withdrawn from the soil. In much of Canada, water-tables rise until freeze-up when most precipitation is in the solid form and little soaks into the ground. During the winter there is a gradual falling of the groundwater levels until spring break-up. Even though frozen ground would seem to seal off the surface and ice and snow would not be expected to provide much water until they melt, evidence is accumulating that some groundwater recharge takes place through frozen ground and beneath the snow.

Groundwater Flow Systems

Darcy's formula for the flow of fluids through porous media may be stated as: the quantity of water flowing through a given cross-sectional area=the permeability × the hydraulic gradient. In groundwater flow systems the permeability is a fixed property of the geological materials and consequently both the hydraulic gradient and the quantity of water flowing are dependent on the geology.

Water moves through geological materials, through not only intergranular spaces but also open fractures that may vary in size from minute to actual solution cavities many feet in extent. Movement through intergranular spaces in porous materials is diffuse and fairly uniform but the most permeable materials, sands and gravels, are commonly laid down by running water, and consequently have directional properties; the permeability in one direction may be many times that in another. Normally the horizontal components are greater than the vertical but even these may vary greatly in direction and between individual sand layers in a single sand deposit. These layers may be only inches or a fraction of an inch thick and still have different directions of maximum permeability. In other geological materials, permeability may be due largely to fractures and bedding and the direction of maximum permeability may be in any direction; in layered rocks it is commonly horizontal but in some fractured and highly jointed rocks it may be vertical.

It is the varying permeabilities and the direction of greatest permeability that give rise to the seemingly anomalous occurrences of groundwater which have led to belief in mysterious underground lakes and streams. Each case depends on its own local conditions but a few examples may illustrate this. In an area with deep permeable materials and a deep water-table, a layer of low-permeability material above the water-table can impede the downward flow of recharge water and in effect have its

own small isolated water-table above it. These are known as perched water-tables or perched aquifers (Fig. 3). In some places, deep valleys have been eroded and filled with very porous material and later deeply covered by less permeable material. These may be highly productive aquifers or even give rise to lines of flowing artesian wells, and, because of the relative scarcity of water on either side, may be interpreted as underground streams, which they are, though the flow is through permeable materials and not through an actual channel-way as in a surface stream. In bedrock areas where the flow is largely through fractures, a well may penetrate deeply without intersecting a fracture of any appreciable size, whereas a well a few feet away may intersect several good fractures and produce considerable quantities of water.

Groundwater flow systems may vary in scale from those related to a single small isolated pond or swampy depression to systems many hundreds of miles in extent, as indicated by oil well data, or even to very deep flow systems of continental size (Meyboom, 1963). Small local flow systems can be variable and a small depression that is a discharge area at one time of the year may be a recharge area at another. The various sizes of flow systems are interconnected much as eddies and currents in a river, but in general the larger the flow system the more constant its flow. It is these larger flow systems with their comparatively constant flow that provide the water that keeps rivers and streams flowing during long dry periods. Thus, the low flow of rivers gives a close measure of the amount of water travelling through related flow systems.

There is much evidence for groundwater flow systems, but only the broader principles used in studying these are considered here as the details can be found in papers on particular flow systems (Hubbert, 1953; Meyboom, 1963; Tóth, 1963). Some of the evidence is direct by measurement or by visual observation but much is indirect depending on the effect of water on other things.

The most direct evidence is the presence of a spring where water can actually be seen flowing from the ground. Another direct method is to plot water pressures at various points in the ground by measuring wells and thus obtaining a direct measure of the hydraulic gradient and consequently the direction of flow. Tracers of various types—dyes, chemicals, or radioactive materials—can be injected into the groundwater flow system either naturally or artificially, and by detecting their appearance in wells or discharge areas the direction and approximate speed of flow can be determined.

Indirect methods are much more varied but largely depend on the chemical composition of water. As water is a very active solvent, it is continually dissolving and precipitating material as it travels through the earth. On a large scale and regardless of the materials through which it travels, it starts out as a bicarbonate water as deposited by precipitation, and changes through a sequence of bicarbonate plus chloride to chloride plus bicarbonate to chloride plus sulphate or sulphate plus chloride finally to a predominantly chloride water approaching the composition of sea water (Chebotarev, 1955). This full metamorphic sequence is not always present in any flow system but has been found a very useful indicator of flow systems in several areas of Canada (Charron, 1960, 1963; Tremblay, 1961). The composition of water also depends on the materials through which it flows as water in limestones and dolomites will be high in carbonate and that in gypsum-bearing rocks high in sulphate, while in clean sands and gravels very little material will be dissolved.

Not enough is known about the rate of change of dissolved material in groundwater but evidence that it can be rapid is accumulating, as changes in composition of groundwater depend closely on the containing materials even though the flow system may cross the boundary between them. However, it may be stated in general that areas of recharge will be of better quality water than areas of discharge.

These changes in composition are reflected in plant growth, and consequently an understanding of plant ecology can lead to an understanding of the groundwater flow system. In many areas, for example, salt-tolerant plants indicate areas of discharge from long flow systems while non-salt-tolerant plants are found in recharge areas or over areas of lateral flow.

Plants may also be divided into phreatophytic and nonphreatophytic types. Phreatophytic plants depend on a continuous supply of moisture and normally grow in places where their roots can reach the water-table. Nonphreatophytic plants depend largely on soil moisture and are able to survive considerable periods with little or no moisture. Thus in regions where soil moisture is scarce phreatophytic plants grow mainly in discharge areas and nonphreatophytic plants in recharge areas.

Geological Materials

Geological materials may be divided into three broad groups for hydrogeological purposes: surficial materials, sedimentary rocks, and crystalline rocks.

Surficial materials contain many of the best local aquifers and as they are near surface are most economically tapped. These deposits comprise all the unconsolidated material lying on top of bedrock, and include the broad size range of glacial and alluvial materials varying from the finest grained, most impermeable clays through all gradations to coarse, clean sands and gravels; similar materials laid down in lakes and seas; wind-blown deposits; and rubble that has fallen down the sides of hills.

Sedimentary rocks because of their great volume provide the largest aquifers, but because they are present to very great depths they are commonly not tapped as easily nor economically as the surficial deposits. They consist of all the consolidated rocks laid down by sedimentary or chemical processes whether they are flat lying or folded gently or steeply. They grade into crystalline rocks in some areas due to metamorphic processes but in most areas overlie the crystalline rocks.

Crystalline rocks are with few exceptions the poorest aquifers because of their low porosity and permeability, and because they are the hardest rocks and consequently much more expensive to drill. They include all igneous and metamorphic rocks that have a crystalline texture.

Surficial Materials

Unconsolidated surficial materials cover all of Canada except for those areas where bedrock actually outcrops. These are the youngest geological materials and some are being deposited at present. Their surface as seen today over great areas is

essentially the same as it was when it was deposited because there has been little time for widespread erosion. Canada was almost entirely covered by Pleistocene glaciers; thus the surficial materials are predominantly of glacial origin.

Glacial Materials

Glacial materials form a complete sequence ranging from those deposited directly from the ice to those deposited directly from running and standing meltwater. Consequently they consist of a complete gradation of sizes from rock flour and clay through sand and gravel to huge blocks, and a complete gradation of shapes from angular to perfectly rounded fragments. They generally consist predominantly of fragments of the underlying bedrock, though there may be appreciable quantities of material that has been transported many tens or even hundred of miles (Stalker, 1956). Because the processes of deposition change rapidly the characteristic of most glacial deposits is heterogeneity.

The materials with the greatest lateral extent are till or boulder clay and glacial lake clays and silts. Till is plastered over the land beneath the ice, and is composed of a mixture of all sizes of material though the finer clay, silt, and sand sizes are predominant. Till may be structureless or contain layers and lenses of different materials. Many prairie farms depend on sand lenses in till for their water supply. Till sheets commonly are extensive and the thicker ones, 100 to 300 feet and in some places 1,000 feet, may extend for hundreds of miles while the thinner tend to be patchy. Tills conform to the underlying surface though they do mask details. Lake clays and silt are deposited in lakes dammed by ice and may be massive or finely banded (varved). They may also contain beds or lenses of finer material such as the large sand and gravel deposits in Lake Agassiz clays near Winkler, Manitoba. Their extent depends on the size of the lake and some of these were many times larger than the Great Lakes of today. Their thickness depends on the supply of material and the length of time the lake existed and may vary from a few inches to a hundred feet or more. As they were deposited by settling in water, their bottom conforms to the underlying materials and the top surface is extremely flat as in the Regina clay plain.

Ablation materials deposited directly from the margins of the ice or from the upper surface as it melts are unsorted like till, but have not been compacted by the weight of the ice and may have a very irregular surface expression. They commonly mask the underlying topography. They may be thin or hundreds of feet thick as in some moraines in Ontario.

During melting of the glaciers, water in large amounts was present on and around the glaciers, and the deposits left by this water are mixed with and deposited on those originating directly from the ice. Where the water was moving it has sorted the materials and left deposits of sand and gravel while it carried the finer silt and clay away to be deposited in lake basins or estuaries of the sea. Streams flowing through or over the glaciers left their sand beds in the form of eskers and some of these sinuous ridges extend for hundreds of miles. Other outwash deposits are much like stream deposits of today, but because of the large amount of water flowing and of sediment available very large outwash deposits were built in some areas, such as the Great

Sand Hills of Saskatchewan. The shape, thickness, and extent of these deposits depended on the topography of the underlying bedrock or other underlying material, they filled low areas or valleys. These deposits occur today as buried valleys such as the buried Missouri Valley in southern Saskatchewan. Deposits left by moving water are limited to the areas along which water flowed and are narrow and sinuous, except where they spread out in the form of deltas and merge into lake or sea deposits.

Post-glacial Materials

After the glaciers melted, the present drainage system developed and the glacial materials were eroded and reworked by processes that we see in action today. The thickness and extent of these deposits depends on the slope of the stream, the width of the valley, and the supply of material; under favourable conditions they can be many hundreds of feet deep but are always long and narrow. The finer materials settle in lakes or in the sea and form extensive mud deposits. Large areas of sand left by the glaciers and glacial waters have been acted on by high winds to form sand dunes, and in some areas these have spread sand well beyond its original area of deposition. Finer dust is also picked up by the wind and deposited as loess but these deposits are not important in Canada. On any steep slope materials can be loosened by frost action or other processes and moved downhill by gravity to form the rubble that is found in talus slopes. In areas of high relief these rubble masses may be of considerable extent but always lie along the base of a mountain wall.

Hydrogeology of Surficial Materials

The hydrological properties of surficial materials depend largely on their intergranular porosity and permeability which in turn depend on the size, sorting, and degree of compaction of the material. In general, materials sorted and deposited by running water will be the best aquifers and groundwater flow will be concentrated through them.

Very fine silt or clay or other sediments containing large proportions of silt and clay have almost zero permeability and consequently make poor aquifers although they have a high porosity. However, some of these, although in themselves poor aquifers, do have structures that allow them to be used for smaller water supplies. Due to drying out and possibly to other causes, such as volume changes resulting from decrease in overburden pressure, and earth tides, systems of narrow cracks or joints develop. These openings if closely spaced, make the silts and clays usable for small domestic water supplies. These joints extend to depths of tens of feet and there is some evidence that they may extend the full depth of these materials.

Sedimentary Rocks

The chief characteristic of sedimentary rocks is that they occur in comparatively thin sheets of great lateral extent, this quality is imposed on them by their mode of deposition. Most sedimentary rocks are deposited in large bodies of water and the rock-forming materials are the products of erosion of the land adjoining the water.

Thus, whether deposited physically or chemically the materials blanket the surface of the bottom of the body of water. Very slight changes in the provenance of the material cause changes in the sediments being deposited and thus create the characteristic flat-lying bedding in the sediment. The thickness of the beds depends on the rate of supply of material and the duration of uniform conditions of sedimentation. Beds may vary from extremely thin to several tens or even hundreds of feet in thickness though beds of a few feet or tens of feet are the most common. The total thickness of a whole sedimentary sequence may vary from a few feet to tens of thousands of feet and the lateral extent may range from a few miles to many thousands of miles.

Material transported mechanically by water or air is sorted according to the weight and size of the particles because of their different settling rates. Thus coarse, more permeable material commonly occurs near the bottom of a continuous sequence, closer to the source, and sediments deposited at the same time may be permeable near the source and impermeable away from it (facies change). In general, finer material is more easily transported and fine-grained sediments may be expected to be more extensive than coarse ones. However, due to the continuous nature of geological processes even coarse-grained sediments can extend many hundreds of miles.

Materials carried in solution and deposited by chemical processes may be carried for very long distances and thus are commonly continuous for hundreds or even thousands of miles. Sediments formed by organic processes may be of great lateral extent if they are deposited by organisms floating in the water, or may be confined to comparatively narrow limits if they are built up from the sea floor and require a limited range of living conditions for the organisms that build them, as for example reefs. Sediments deposited by wind are normally fine-grained sands or silts and not as extensive and continuous as water-lain sediments.

The processes of erosion and sedimentation are slow and continuous but change in time and place so that erosional surfaces can be buried by younger sediments. These buried erosion surfaces may be very extensive but as they have a topography like the present surface and vary in thickness quite erratically they may be very irregular. Due to changing conditions all these processes may be repeated many times and be mixed so that the sedimentary succession can have a great variety of rock types. If one understands the methods of deposition of sedimentary rocks and studies their history, it is possible to make reliable predictions of their extent and properties and this is essential to understanding their hydrogeology.

Hydrogeology of Sedimentary Rocks

The hydrogeology of sedimentary rocks depends both on their primary physical features and on secondary features formed after their deposition. The more important primary features are the flat or nearly flat beds, the grain size as related to its sorting and distribution, the degree of cementation, the position in the stratigraphic column, and facies changes.

The flat or nearly flat beds extend long distances and allow for extensive porous and permeable zones whose thickness depends on the thickness of the bedding. Their ability to store and transmit water depends on the size, shape, and sorting of their

constituent particles. Those formed of particle sizes from coarse silt up have high porosity and permeability if the particles are well rounded and of uniform size, because this leaves well-connected openings between grains as in a bag of marbles. Conversely if the particles are angular and a mixture of sizes, they fit together, leave poorly connected openings, and thus have low porosity and permeability. There are of course all gradations between the two extremes and consequently all combinations of porosity and permeability. The smaller sizes from fine silt down may have high porosity but because of the minute openings have very low permeability, and while there may be lots of water in storage it cannot be extracted. Porosity and permeability also depend on the degree of cementation. Grains may be cemented at their points of contact thus reducing the permeability greatly but the porosity only slightly, or the cement may fill the openings almost entirely thus reducing both porosity and permeability almost to zero as in many chemically deposited sediments. The position in the stratigraphic column may be important as permeable rocks bounded above and below by impermeable ones may either be sealed off from recharge or have the water held under pressure and be artesian aquifers. Of equal importance are facies changes which may result in a permeable bed in one area being impermeable in another.

With all these possible variations in sedimentary rocks, a thorough knowledge of the sedimentation processes and history is required before the groundwater flow systems can be evaluated and predicted. Of possibly even greater importance hydrogeologically are features imposed on sediments after their deposition. These secondary features are openings of various types from small cracks to large caverns and they can impart varying degrees of porosity and permeability to any type of rock regardless of its original composition. They are all due to changes in geological environment, varying in severity.

Joints are narrow fractures that are due to relief of stresses developed during slow raising or lowering of the earth's crust, to volume changes during consolidation, to continuous minor movements due to earth tides, to relief of stress due to irregular erosional unloading, or to more violent forces of folding and mountain building. They are normally straight fractures of a few inches to many hundreds of feet in extent but commonly are not continuous for more than a few tens of feet. They occur in subparallel sets of various attitudes and in most areas there are at least three intersecting sets and sometimes many more. In sediments, they commonly form perpendicular to the bedding and, depending on the characteristics of the rock, they may be confined to a single bed or be continuous through several beds. If the sediments are folded the joints are accentuated and folded with them so that they still remain perpendicular to the beds, even though the beds may be tipped on edge or overturned. In folded areas, other forms of fractures may form at angles to the bedding but the important thing is that there are almost universal fracture systems connecting the bedding planes and the bedding planes themselves are a form of fracture through which water moves.

Faults are fractures in rock along which the rocks on one side have moved relative to the rocks on the other. The amount of movement can vary from a fraction of an

inch to many miles and the movement can be in any direction within the plane of the fault, but the amount and direction of movement are hydrogeologically important only because they bring rocks of different hydrogeological properties together and also because considerable localized movement provides more opportunity for severe fracturing of the rocks. Faults vary in length from a few feet to hundreds of miles and may be straight though more commonly are sinuous. They may be steep or flat lying and be single isolated features, or branch into numerous faults, or even occur in sets like joints, into which in fact they grade, as a fault with no movement is a joint. Fault fractures may vary from narrow fractures completely sealed by finely ground rock or fault gouge to wide bands of shattered rock, in some instances in the order of a mile wide, and highly porous and permeable. All gradations between these two extremes exist and conditions may change from place to place along a fault. Also, chemically deposited material may seal them. Thus while faults can be traced with considerable accuracy, their hydrological properties are unpredictable and they may be productive aquifers or barriers to water movement.

Movement on a fault may bring an aquifer against an aquiclude, and if the fault is at all permeable then water under pressure in the aquifer may escape to surface along the fault. Thus some faults are shown by lines of springs as along the Gloucester Fault between Ottawa and Cornwall. Others as the subparallel sets of faults in the Foothills region of western Alberta are barriers, and groundwater in the miles-wide slices of rock between them acts independently of that in adjoining slices.

Folding of sedimentary rocks can change their hydrological characteristics not only by tipping up the flat-lying beds to varying degrees but also by increasing the degree of fracturing. The degree of fracturing depends on the rock material as well as its relation to adjoining rocks. Some rocks are brittle and will fracture when folded while others are soft and will flow without creating any fracture openings. The degree of brittleness is not necessarily connected with any original hydrological properties of the rock, but is important in relation to the overlying and underlying beds, as it is possible for a rock that initially had little or no permeability, but was brittle and lay between two more permeable but soft rocks, to be fractured so completely during folding that its fracture permeability becomes greater than the natural permeability of the two adjoining rocks. In brittle rocks that are sharply folded, extensive zones of fractured rock may be formed along the axes of the folds and be productive aquifers.

Water moving through any of these various forms of fractures can dissolve material and over a sufficient period of time create interconnected caverns of considerable extent. Such conditions are known in many areas of Canada, one of the most extensive of which is in the interlake area of Manitoba. Wherever weathering processes have opportunity to work on sediments for long periods of time, various chemical and physical processes act upon them and result in increased porosity of the upper few feet or tens of feet of rock. These weathered zones may later be buried by other rocks or surficial materials and be productive aquifers. Only an understanding of the history of the geology of an area can enable these weathered zones to be predicted and utilized.

Crystalline Rocks

Crystalline rocks are formed by a wide variety of processes usually involving high temperature and pressure. They include an even wider variety of rock types but all are composed of tightly fitting and interlocking crystalline mineral grains. Crystalline rock masses may be of almost any shape or size varying from narrow tabular dykes, to massive uniform batholiths, to areas of mixed metamorphic rocks. Actual sizes vary from a fraction of an inch to hundreds of miles of similar rock, and shapes may vary from extremely regular and predictable to completely irregular and unpredictable, the latter of which is more common. Their shape and extent can only be determined for any area by actual field mapping.

The greater part of the surface of Canada is underlain by crystalline rocks and even areas of sedimentary rocks are underlain at some depth by crystalline ones. Consequently they are the only aquifers in large parts of the country where surficial materials are thin and sediments lacking. Because of their great extent they do contain vast amounts of groundwater and many areas depend entirely on them for their water supply. Some idea of the amount of water present is given by the large volumes of water that are pumped continuously from mines (Table III).

Hydrogeology of Crystalline Rocks

Crystalline rocks display many of the secondary features described for sedimentary rocks but because of their generally massive structure there are some differences. Primary features in igneous rocks are of little hydrological significance. Metamorphic rocks by definition are in themselves secondary and their hydrological features are similar to those of igneous rocks with the exception that they may have banding that may result in hydrogeological characteristics similar to sedimentary bedding.

Joints formed by cooling and shrinking of igneous and metamorphic rocks as well as by all the forces described for sedimentary rocks are the chief sources of porosity and permeability in crystalline rocks. They are more continuous than those in sedimentary rocks and sometimes extend for many miles though normally they are consistent over comparatively short distances. Quarry and mining operations show that the most common sets are two or three steeply dipping sets and one almost parallel with the earth's surface, and that the steeper ones in particular are present at the greatest depths yet mined (about two miles), while the flat-lying ones are most numerous near the surface and become much more widely spaced below a few hundred feet. In general, all joints are more numerous and open near the surface, and below a few hundred feet tend to be tight though considerable flows of water have been encountered from joints in mines at depths of several thousand feet. Joints are present in all but a few extremely massive crystalline rocks and are most abundant in metamorphic rocks as the latter have been subjected to the greatest amount of disturbance. It is the almost universal presence of joints and the interconnection of the various sets that makes it possible to use crystalline rocks as aquifers; they are widely used throughout Canada though rarely for large supplies.

No quantitative studies have been done on groundwater in crystalline rocks in

Table III

Selected List[1] *of Canadian Mines Showing Water Pumped Daily*

(Figures are approximate and do not separate water introduced for mining purposes but indicate an order of volume)

Mine	Location	Ore Mined	Depth (ft)	Water Pumped (gpd)
Nova Scotia				
MacBean	Thorburn	coal	1,275	600,000
Greener mine	Sydney Mines	coal	280	575,000
Walton mine	Walton	barite	970	2,700,000
New Brunswick				
Wedge	Newcastle	copper–zinc	1,150	150,000
Heath Steele	Newcastle	copper–zinc	400	720,000
Quebec				
Gaspé	Murdochville	copper	700	1,100,000 to 6,000,000
East Malartic	Norrie	gold	4,950	500,000
Johnsons	Thetford Mines	asbestos	1,200	115,000
Kilmar	Kilmar	magnesite	850	1,300,000
Newfoundland				
Rothermere	Buchans	lead–zinc	2,500	20,000
Tilt Cove	Tilt Cove	copper	2,700	600,000
Ontario				
Langis Silver	New Liskeard	silver	400	200,000
No. 2 mine	Caledonia	gypsum	75	75,000
Madsen	Madsen	gold	4,075	109,000
Errington	Atikokan	iron	1,480	2,100,000
Frood	Sudbury	nickel–copper	3,600	300,000
Lake Shore	Kirkland Lake	gold	8,075	200,000
Manitoba				
San Antonio	Bissett	gold	4,930	7,500
Amaranth	Amaranth	gypsum	130	1,000
Flin Flon	Flin Flon	copper–zinc	4,000	616,000
Saskatchewan				
Beaverlodge	Uranium City	uranium	4,000	280,000
Alberta				
Atlas	East Coulee	coal	400	86,000
British Columbia				
Bluebell	Riondel	silver–lead–zinc	1,110	5,900,000
Britannia	Britannia Beach	copper–zinc	4,000	up to 6,000,000
Northwest Territories				
Con-Rycon	Yellowknife	gold	3,500	470,000
Yukon Territory				
Hector and Calumet	Elsa	silver–lead–zinc	1,300	50,000

[1]From list compiled by A. Dubnie, Mineral Economics Section, Mineral Resources Division, Department of Mines and Technical Surveys, in 1962.

Canada, but studies of similar rocks in other countries indicate generally that banded crystalline rocks give more water than massive rocks and that optimum depths for domestic wells are less than 200 feet, for larger production wells less than 600 feet (Bryn, 1961; Davies, 1964). Faults in crystalline rocks, as in sedimentary rocks, may vary greatly in hydrological properties but steeply dipping permeable ones can carry water to great depths and at great pressure. A mine at Yellowknife, Northwest Territories, encountered such a fracture at a depth of 2,300 feet, in which the water was at a pressure that indicated that it must be an open fracture filled with water almost to surface. Tunnelling operations involved in the Aluminum Company of Canada's developments at Kitimat, British Columbia, encountered zones thousands of feet long in which the rock was so completely fractured by faults and joints that tunnels had to be lined with steel and concrete supports not only to hold back large flows of water but to support rock that was being loosened by this water (Mathias, 1954). Thus, under favourable conditions faults and associated fracture zones can be major aquifers even though they are not as widespread as joints and do not have the same degree of interconnection.

Because porosity and permeability in crystalline rocks depend on interconnected fractures, the production of a well depends mainly on the number of fractures intersected and only to a minor extent on its diameter. Also, because of the erratic distribution of fractures it is possible to have a dry hole only a few feet from a productive well.

Weathering processes affect crystalline rocks as they do sedimentary ones, but as crystalline rocks are formed mainly of insoluble minerals, fractures are seldom widened into solution channels. Long continued weathering can cause chemical changes in the minerals of crystalline rocks developing clay minerals with accompanying volume change or dissolving out selectively more soluble minerals. Both these processes may leave a porous and permeable surface zone and it is probable that at one time most of the crystalline rocks in Canada were covered by a thick weathered zone that graded up into surficial materials, but during the glacial periods most of this has been scraped off and now only comparatively small pockets, seldom more than a few miles in extent, of this type of material are found unless they are protected by more resistant rocks. An example of this is the Nepean sandstone in the Ottawa area where both the top and bottom are weathered zones and locally are good aquifers, while the main body of sandstone itself is a poor aquifer because it is a well-cemented sandstone, almost a quartzite, whose only permeability is through fractures. Folding in crystalline rocks is of minor hydrological importance though in some banded metamorphic rocks sharp folding can create fracture zones at the crests of folds. A few crystalline rocks do not fall into the general description just given because of their peculiar composition or properties. Crystalline limestone is a soluble rock and fractures within it may be enlarged by moving water to become solution cavities or even caves; in rare cases these may be large water producers.

Fresh volcanic rocks may be very porous and permeable due to fragmental zones and gas cavities formed when they solidified; in some areas, such as central British Columbia, spring lines around volcanic areas indicate that they can transmit considerable water.

Chemistry of Groundwater

Water is a very complex chemical substance and when it becomes part of a groundwater flow system it is in intimate contact with a vast number of minerals with a wide range of chemical properties, and is subjected to an equally wide range of time, temperature, and pressure conditions. Obviously it would be impossible to explain all the interactions of water and its geological framework so the following only attempts to explain in broad terms how the chemistry of water relates to the materials through which it is flowing.

Absolutely pure water is comparatively inactive and water depends for its activity on its temperature and pressure, its acidity or alkalinity, its pH, and dissolved ions. Precipitation is originally distilled water but it picks up oxygen and carbon dioxide while falling through the atmosphere. Thus, recharged water is both oxidizing and acid when it infiltrates through the zone of aeration above the zone of saturation and many complex chemical reactions take place. At the start of a flow system, water is most active and changes take place rapidly until it approaches chemical equilibrium with its surroundings. Once equilibrium is approached water reacts with its surroundings slowly and the chemistry of the flow system is comparatively constant. However, if the flow system crosses into a contrasting formation, the water may suddenly be chemically active relative to the new formation and a series of changes are initiated that proceed until it again approaches equilibrium with its surroundings. For this reason, changes in the chemistry of groundwater are normally rapid at the boundaries between formations, and the chemistry of the water is closely related to the enclosing geological materials. Rocks composed of comparatively insoluble materials, such as quartz and feldspar, have little effect on the chemistry of water flowing through them, so that the water retains any characteristics that it had on entering them. Rocks in this class consist of most crystalline rocks and the surficial deposits and sediments formed from them by erosive processes that have eliminated all but the most insoluble materials or clean well sorted sands, gravels, sandstones, and conglomerates.

Carbonate rocks such as limestones, dolomites, and crystalline limestones dissolve rapidly in any acid water and will yield water high in calcium, magnesium, and carbonate. The amount of these materials in solution depends on the acidity of the water and as this acidity is used up in dissolving the rock there will be little more solution once the water becomes neutral or alkaline. Rocks deposited by evaporation of shallow seas, called evaporites, consist largely of sodium, potassium, and magnesium sulphates and chlorides and are highly soluble so that water flowing through or near such rocks will be very high in these materials. Shales and some crystalline rocks may contain large quantities of soluble iron sulphides and water flowing through them can contain much iron and hydrogen sulphide. Clay minerals have peculiar properties of ion exchange and also commonly contain many calcium and magnesium minerals. The ion exchange property can exchange sodium for calcium and magnesium and consequently rocks containing clay minerals can produce water very high in sodium.

Geological materials and rocks can be mixed in almost any possible proportion and consequently groundwater can have any mixture of dissolved contents and the

amount of dissolved material can range from a few parts per million to several hundreds of thousands of parts per million. For comparison sea water contains about 35,000 parts per million dissolved solids. In a very general way, the deeper the groundwater flow system, the more dissolved material, because such flow systems are in contact with the rocks for longer times at higher temperatures and pressures.

There is also the problem of man-made pollution of groundwater with almost any type of chemical. The presence of a high nitrate content is usually a danger sign indicating artificial pollution. The health problems of pollution and the presence of fluorides or other dissolved minerals are a major study in themselves, and as they fall within the realm of public health will not be considered here.

Data on the chemistry of groundwater found in the various hydrogeological regions of Canada are given in the chapters describing each of these regions.

Hydrogeological Regions

If studies are conducted of typical areas within regions where hydrogeological conditions are similar, the knowledge gained may be used to estimate conditions in other areas within the region. By combining broad factors of geology, climate, precipitation, and topography one may divide Canada into six major geological regions. Chapters III to VIII describe groundwater conditions in each of these regions. Some are described in much more detail than others, and this is a direct reflection of both the amount of data available and the amount of work that has been done to answer problems in each; it also indicates where future work is required.

Figure 7 (*in pocket*) shows the boundaries of the six regions. It should be remembered that these boundaries do not represent sudden changes but are zones of change. As knowledge increases, it will be possible to subdivide each region into more useful units.

The Appalachian Hydrogeological Region is characterized by a mixture of crystalline and sedimentary rocks, mostly folded but with some areas of flat-lying sediments and comparatively thin surficial materials except locally. The climate is humid continental, and annual total precipitation ranges from 30 to 50 inches with the maximum along the Atlantic coast. Topographically, this region is an upland sloping gently to the southeast, dissected by valleys and broken by broader lowland areas developed on belts of weak rocks. Relief in the Gaspé region exceeds 2,000 feet but elsewhere it is seldom more than 1,000 feet and commonly only a few hundred feet.

The St. Lawrence Lowlands Hydrogeological Region is underlain by unfolded sedimentary rocks overlain, in most places, by a fairly thick cover of surficial materials. The climate is humid continental, with 30 to 40 inches of total precipitation per year. Local relief is generally low and seldom exceeds 200 feet.

The Canadian Shield Hydrogeological Region consists almost entirely of mixed crystalline rocks with only minor areas of sedimentary rock. Surficial materials are distributed irregularly with little or none on the upland areas and thick deposits largely confined to the valleys. The climate is humid continental; total annual pre-

cipitation varies from 20 inches in the western part to as much as 40 inches in some of the eastern parts. The topography is extremely rugged, though relief is seldom more than a few hundred feet except for a few areas where it exceeds 1,000 feet.

The Interior Plains Hydrogeological Region is underlain by a thick sequence of unfolded sedimentary rocks except for a folded belt extending along the western edge. It is almost entirely covered by a thick layer of surficial deposits, and bedrock outcrops are seldom seen except along the deeply incised river valleys. The climate is sub-humid continental except for the area between Moose Jaw and Lethbridge and as far north as latitude 52°N where it is arid to semi-arid. Total precipitation varies from 10 to 20 inches annually and exceeds 20 inches only at the extreme western and eastern limits. The elevation of the land falls from about 4,000 feet on the western edge of the region to 500 feet at the eastern edge. Although the name "plain" suggests a level tableland, most of the surface is rolling and sometimes hummocky with relief of 100 to 200 feet except for the flat glacial lake basins as at Regina and Winnipeg. In addition, many irregularities of topography are present both as deeply dissected river valleys that may be as deep as 400 feet and occupy steep-walled valleys 1 to 2 miles across, and as prominent flat topped hills standing 200 to 600 feet above the general level.

The Cordilleran Hydrogeological Region is largely underlain by crystalline rocks and steeply folded sedimentary rocks with some flat-lying sedimentary rocks in the central plateau and coastal area. Surficial deposits are confined to the valleys and reach thicknesses of 700 to 800 feet and possibly much more. The climate varies from marine west coast to semi-arid and is extremely variable depending on relations to mountain masses. Total annual precipitation varies from less than 10 inches in some parts of the high, deeply dissected central plateau to 40 inches in the eastern mountain systems and more than 100 on the coastal side of the western mountain systems. Relief in the mountain systems is many thousands of feet. In the central plateau, while the upland areas show relief of only a few hundreds of feet, numerous valleys have been cut several thousand feet below this surface. These extremes of topography and precipitation make the main hydrogeological characteristic of this region one of extreme variability.

The Northern Hydrogeological Region is vast and hydrogeological data are so scattered that it has not been considered worthwhile to subdivide at this stage of our knowledge. From Great Bear and Great Slave Lakes eastward the region is underlain by crystalline rocks except for an area southwest of Hudson and James Bays where there is a cover of unfolded sedimentary rocks. Between this crystalline rock area and the mountains to the west lies a broad belt of mostly unfolded sedimentary rocks, and from the mountains to the western limit of the region is a complex area of folded sedimentary rocks mixed with crystalline rocks. Surficial materials vary greatly in their extent and thickness, and along the Mackenzie River and in the Yukon reach their greatest thickness in the valley bottoms. The climate in the crystalline rock areas is mostly sub-Arctic, in the unfolded sedimentary rock areas is semi-arid, and in the mountainous areas is mostly sub-Arctic, with tundra in the northern parts of all areas. Total annual precipitation is less than 20 inches everywhere north and west of a line

drawn roughly from the south end of Ungava Bay to the south end of Lake Winnipeg and is less than 10 inches from Great Slave Lake north to the Arctic coast and east to Hudson Bay. Southeast of the Ungava Bay–Lake Winnipeg line it increases to more than 30 inches near the Canadian Shield region. Relief is varied but east of the mountainous area, the Yukon, it is commonly only a few hundred feet and seldom exceeds 1,000 feet except in the extreme north and east. In the mountainous areas to the west relief is similar to the Cordilleran region.

Basin Studies

The study of groundwater is the study of a dynamic system from its source in the area of recharge through its flow system to its end in the areas of discharge. The eventual aim of all studies is to gain a full knowledge of the quantity and quality of water entering this flow system, where it flows, what chemical changes take place, where it discharges, and what fluctuations can be expected from both natural and artificial causes. Complete flow systems are best studied in their entirety in order to minimize outside interfering factors. Also detailed studies of particular groundwater problems are a study of part of a flow system.

By definition a drainage basin is an area encompassing the surface flow-system, and investigations show that to a certain depth it generally coincides closely with the associated groundwater flow system. For this reason drainage basins have been chosen as the most practical unit to use in studying groundwater flow systems. As knowledge increases and it becomes possible to more closely define groundwater flow systems, particularly at greater depth, studies will be directed more towards flow systems not defined by detailed surface topography but by general surface elevations and the hydrological properties of the rocks through which the water flows.

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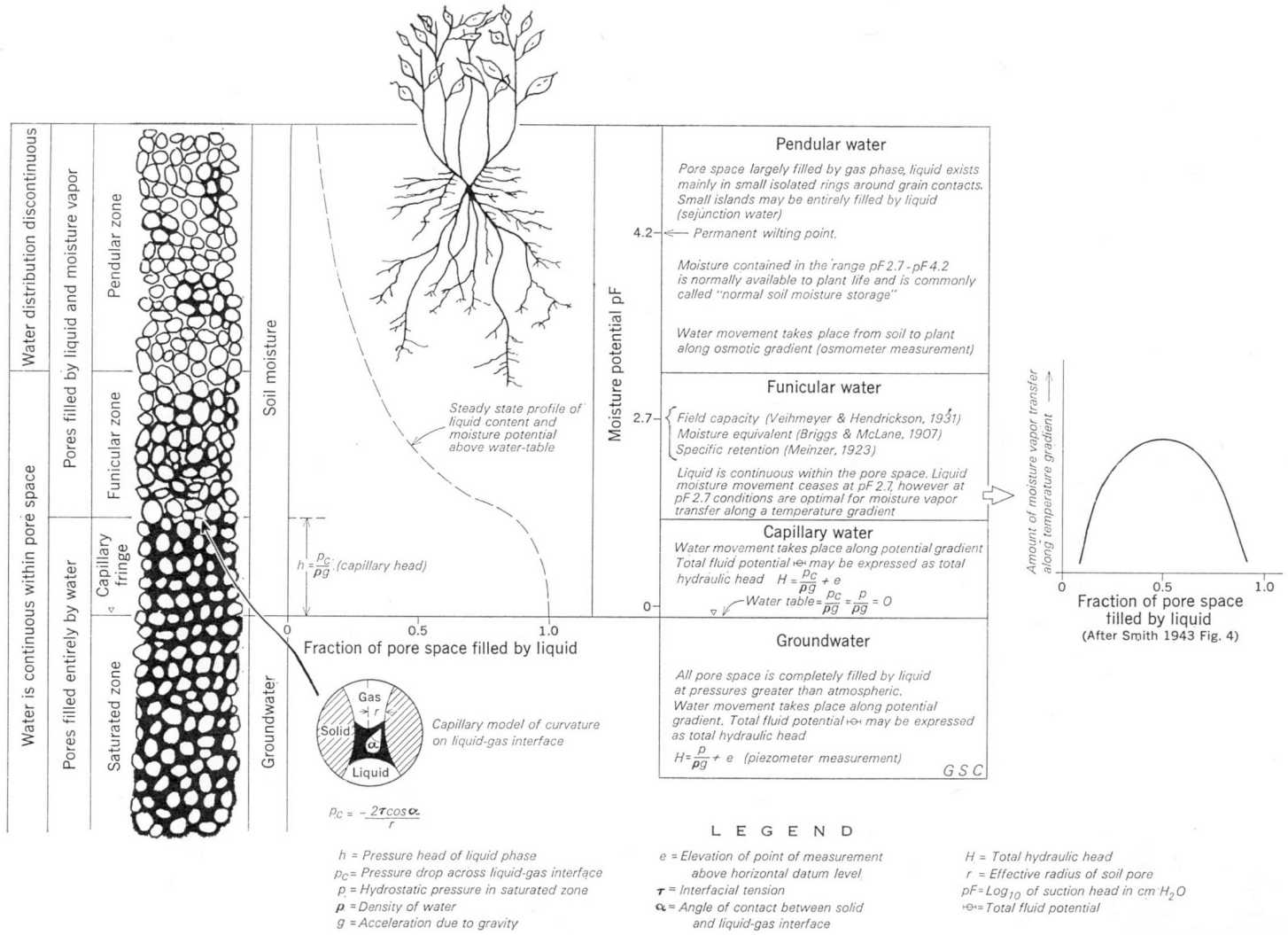


FIGURE 8. Schematic presentation of liquid occurrence in a porous medium and some empirical coefficients pertaining to certain energy conditions.

Chapter II

HYDROGEOLOGY

by Peter Meyboom

Hydrogeology is the science dealing with the character, source, and mode of occurrence of underground water. As such, it is an integral part of geology, the science that deals with the origin and the state of the earth. Many geological phenomena can be understood better with a proper understanding of the groundwater processes that are involved, but, conversely, there is no part in the study of groundwater that can be explained without the understanding of geology. This means that hydrogeology can be used either as an auxiliary to geology, or that it may be used in a wider sense as the study of part of our natural environment. The following outline of principles and methods clearly reflects this dual nature of the subject.

The fundamental working hypothesis of either approach is the continuity of the hydrologic cycle. The prime concern of our investigations is not whether this continuity is economically favourable, but rather how this continuity manifests itself, how it influences other natural phenomena, and how it can be used advantageously in studying nature. Thus, by aiming at a more fundamental understanding of nature, hydrogeology reaches far beyond answering one particular question about our social well-being.

Part 1. Principles of Hydrogeology

Terminology

The Four Stages of Versluys

The classification of the occurrence of fluids in porous media that was made by Versluys (1916) distinguishes among four stages of moisture occurrence, depending on its relative continuity (Fig. 8). In the *pendular stage* the pore space is largely filled by vapour. Liquid exists only in small isolated rings around the grain contacts, which are the first regions where the force interaction between liquid molecules can dominate over the force interaction between liquid and solid molecules (Stallman, 1964). After Versluys, Stallman (1964) defined liquid occurrence as being in the *pendular stage* at that moisture content at which the pendular rings begin to coalesce.

The *funicular stage* is characterized by a moisture content whereby the pendular rings coalesce to such an extent that the liquid films become continuous throughout the pore space, thus entirely enclosing the vapour phase. Versluys (1916) reported a funicular zone of 330 cm in dune sand.

In the *capillary stage* all pore spaces are occupied by liquid, but the liquid pressure within the pores is less than the total pressure outside the system. The capillary stage is brought about by the molecular forces that are responsible for surface energy. In the very narrow pores, water rises similarly as it would rise in a capillary tube, simply because the adhesive forces between the molecules of the liquid and the molecules of the solid exceed the cohesive forces between the molecules of the liquid. Capillary rise above the saturated zone is determined mainly by the effective pore size of the material and varies theoretically from 0.5 foot in fine gravel to more than 150 feet in clay. Under field conditions, however, the capillary rise seldom exceeds 5 feet.

The capillary stage ends at the *phreatic surface* (water-table), at which the atmospheric pressure is in equilibrium with the hydrostatic pressure. All pores beneath the phreatic surface are completely filled with liquid under hydrostatic pressure, called groundwater; this is in contrast to the moisture above the water-table, which is collectively referred to as soil moisture.

Some authors recognize, in addition to those mentioned above, two other stages: the *structural stage*, in the dry state of the medium, which includes the water molecules that form part of the mineral structure of the porous medium; and the *adsorbed stage*, in which the force interaction between liquid molecules is everywhere dominated by the force interaction between liquid and solid molecules.

The Limits of Groundwater

Under natural conditions, groundwater fills the pores and fractures of all rocks. As the temperature of the earth increases with increasing depth at an average rate of 0.02°F/ft, groundwater will be warmer with increasing depth. Spicer (1936) reports that an extrapolation of temperature gradients in a large number of different areas in the United States of America showed that the normal boiling point of water was found at a depth of less than 7,000 feet in a third of the cases and deeper than 10,000 feet in another third.

However, groundwater is also subjected to an increase in hydrostatic pressure of approximately 0.45 psi/ft. The combined effect of temperature and pressure is that groundwater can exist in the liquid phase up to the critical temperature of 374°C. Above this temperature, only the vapour phase remains and no further increase in pressure will liquefy any part of the component. According to Levorsen (1954), it was calculated that the critical temperature occurs at depths consistently greater than 30,000 feet. Hot water vapours at such depths belong to the hydrothermal phase of rock metamorphism (Palm, 1960, Fig. 3) and fall outside the realm of hydrogeology.

The lower limit of groundwater, therefore, has to be taken at the depth of the critical temperature of water, whereas the upper limit is the water-table.

Energy Conditions of Single Phase and Multiphase Fluids in Porous Media

Groundwater Potential

In a recent review of fluids in porous media Stallman (1964, p. 44) stated that in order to study the occurrence and behaviour of fluids in porous media in a rational manner,

> . . . fluid and media characteristics must be expressed in terms of energy dissipation or energy states resulting from fluid movement or state of fluid occurrence.

Energy relations of fluids in the saturated zone have been studied extensively by Hubbert (1940, 1953) who formulated the general condition as follows (1953, p. 1958):

> An element of water at any point possesses potential energy with respect to its environment which, when referred to unit mass, we may speak of as its potential, Φ.
>
> The potential (Φ) of water at a given point may be thought of as the amount of work that would be required to transport unit mass of this fluid from some arbitrarily chosen standard position and state to the position and state of the point considered. For the standard state we may take a closed chamber containing water at elevation z_0 and pressure p_0. The final state is that of the point of interest whose elevation is z and whose fluid pressure is p.

The work done in transporting a unit mass of water from the initial system to the final one is composed of three parts: work against gravity, work against pressure, and if the fluid in the final state system is in motion—work to accelerate the fluid from a velocity of zero to that of v.

The mathematical expression of the fluid potential at a point P can be written in its well-known form as:

$$\Phi = gz + \int_{p_0}^{p} \frac{dp}{\rho} + \frac{v^2}{z} \qquad \text{(Hubbert, 1940, eq. 28) (1)}$$

where Φ=total fluid potential
g=acceleration due to gravity
ρ=fluid density
z=elevation
p=pressure
v=velocity.

As the flow velocity of groundwater is generally extremely small, the velocity potential may be ignored, reducing the expression of total fluid potential to:

$$\Phi = gz + \int_{p_0}^{p} \frac{dp}{\rho} \qquad \text{(Hubbert, 1940, eq. 30) (2)}$$

Elaborating on the practical value of this expression Hubbert said:

> Since the choice of the standard datum is arbitrary, it will be convenient to choose sea-level and atmospheric pressure as our reference state and to set:
>
> z_0=zero; p_0=1 atmosphere.

From this, it follows that the water-table (phreatic surface) is the datum level for hydrostatic pressure. The potential at any point in the flow field is now equal to the

amount of work required to transport unit mass of water from sea-level and atmospheric pressure to the elevation and pressure of the point considered. The expression for the total fluid potential then simplifies to:

$$\Phi = gz + \frac{p}{\rho} \qquad \text{(Hubbert, 1953, eq. 4)} \quad (3)$$

where p=gauge pressure at the point under consideration
ρ=fluid density.

Thus, if the fluid potential at point P has to be measured in an open-top manometer (Fig. 9) terminating at the point P, the potential due to fluid pressure is measured by the pressure p exerted by the column of water that can be supported by P.

$$p = h.\rho g, \text{ or, in terms of head of water, } h = \frac{p}{\rho g}.$$

The potential energy due to the elevation is expressed by the elevation of point P, in feet above sea-level (Fig. 9). It follows that the total fluid potential may be expressed in terms of total hydraulic head H, which is the sum of pressure head $\left(\frac{p}{\rho g}\right)$ and elevation head (z):

$$H = z + \frac{p}{\rho g} \quad (4)$$

This is the most important expression of potential energy of groundwater in the saturated zone. It shows that the potential depends only on the elevation of the point of measurement, the fluid density, and the fluid pressure at the point of measurement.

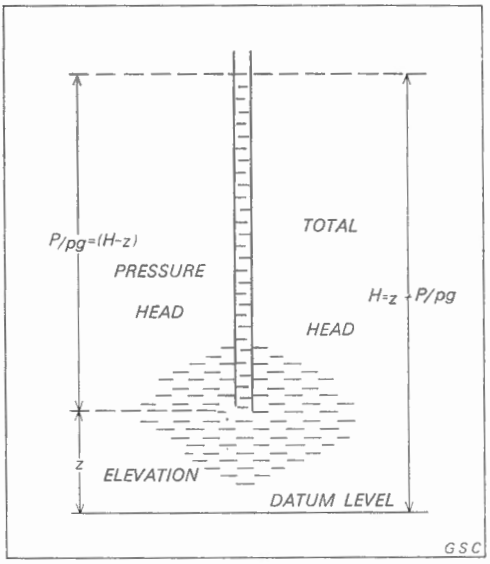


FIGURE 9.
Relation between total head, pressure head, and elevation (Hubbert, 1953).

Measurement of Groundwater Potential

As the total head at any point in the flow medium is an adequate measure of the total fluid potential at that point, an open-top manometer is all that is required for a measuring device. In groundwater studies such instruments are called piezometers. They consist of a small diameter pipe, installed to the depth where the potential has to be determined. Since it is imperative that the total head is measured at one point only, all the formations above the piezometer end are sealed, either by means of a mechanical device (metal petal basket) or by drilling mud and cement (Meyboom, *et al.*, 1966). The column of water in the piezometer is the pressure head at the point of measurement, whereas the elevation head can be determined by conventional surveying.

Moisture Potential

As moisture above the phreatic surface occurs at pressures less than atmospheric, it follows that the unsaturated zone is characterized by negative pressure heads, for which agronomists introduced the term *moisture potential* (sometimes called "soil suction" or "capillary potential"). It can be seen from the following explanation that moisture potential is indeed equivalent to the pressure potential in the saturated zone.

Under static conditions the moisture potential equals

$$\frac{p_c}{\rho g}$$

where p_c is the pressure difference across the liquid-gas interface (Fig. 10), which is expressed by:

$$p_c = -\frac{2\tau \cos \alpha}{r_{ct}} \qquad \text{(Stallman, 1964, eq. 2) (5)}$$

in which τ=interfacial tension

α=angle between solid and gas-liquid interface

r_{ct}=radius of capillary tube[1].

Thus, the moisture potential, being the same as the pressure head of the liquid phase in the unsaturated zone, is expressed by:

$$h = -\frac{2\tau \cos \alpha}{\rho g\, r_{ct}} \qquad \text{(Stallman, 1964, eq. 5) (6)}$$

Both Luthin (1957) and Stallman (1964) stressed that under completely static conditions in the unsaturated zone, the vapour phase must satisfy the following:

$$-H = RT \log p_v/p_0 \qquad (7)$$

in which R=gas constant per gram of the vapour phase

T=temperature in degrees Kelvin

p_v=partial pressure of undersaturated vapour phase

p_0=partial pressure of saturated vapour phase;

or, since H equals $-h$, in terms of pressure head of the liquid phase:

$$h = RT \log p_v/p_0 \qquad \text{(Stallman, 1964, eq. 6) (8)}$$

[1]In agricultural literature r_{ct} is referred to as r, being the "effective radius" of the pore channel in which the liquid rises.

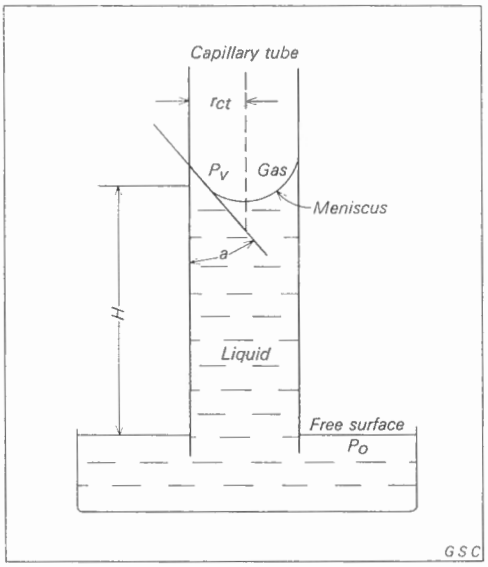


FIGURE 10.
Capillary model of pressure drop across liquid-gas interface (after Stallman, 1964).

Agronomists express moisture potential in cm H_2O suction necessary to extract the moisture from the soil. As this suction may vary from 0 to 10^7 cm water, it is a convenient practice to use the 10_{log} of the water suction in cm, for which Schofield (1935) introduced the measure pF.

It follows from equations (6) and (8) that the pressure head of the liquid phase in the unsaturated zone is susceptible to many external influences that do not affect the pressure head in the saturated zone. The following discussion of some of these effects is largely based on Stallman's comprehensive review of multiphase liquids (Stallman, 1964) and on the excellent introduction to the physics of land drainage by Childs (1957).

Effect of Moisture Content

Childs (1957) described the relation between moisture content and moisture potential most clearly as follows:

> The pore space consists of caverns of various sizes interconnected by channels of diverse widths, as illustrated in Figure 11. A channel which at its narrowest part has an effective radius r_1 can accommodate a gas-liquid interface with curvature of effective radius r_1 or greater, supporting a pressure drop across the interface of p_c^1 or less, r_1 and p_c^1 are related by eq. (5). If the suction increases to a pressure drop across the gas-liquid interface of p_c^2, the interface must be drawn through a channel of radius r_1 until it finds a channel of radius r_2, in which it can come to rest, r_2 and p_c^2 being related by eq. (5). So, as the suction increases, more and more water is extracted, the amount remaining residing in pores with smaller and smaller channels of entry.

A plot of moisture content against moisture potential in these circumstances is of the kind shown in Figure 12.

It is important to note that the same moisture potential can exist at different moisture contents, depending on whether the medium is drying or being wetted. The

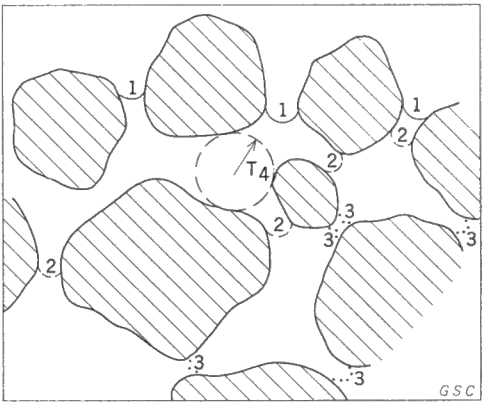


FIGURE 11.
Stages in the change of moisture content of a non-shrinking soil (after Childs, 1957).

hysteresis loops shown in Figures 12 and 13 are explained by the fact that, upon relaxing the suction, the gas-liquid interface in the channel with radius r_2 can not climb back to the channel with radius r_1, because in doing so it must pass a much wider part of the channel, having an effective radius r_4, for example. (Childs, 1957.)

Effect of Temperature

Equations (6) and (8) both contain temperature-dependent variables; consequently, the moisture potential is temperature dependent. The relation between change of temperature and change in moisture potential is given by Stallman (1964, eq. 21):

$$\frac{\partial h}{\partial T} = \frac{h}{\tau}\frac{\partial \tau}{\partial T}$$

where h=pressure head of liquid phase (moisture potential)
T=temperature, degrees Kelvin
τ=interfacial tension,

whereby it is assumed that the moisture content and the moisture density remain constant as the temperature changes.

The effect of temperature indicates that an unsaturated porous medium at a given moisture potential holds less water at a higher than at a lower temperature. Under field conditions, this will manifest itself as slight water-table fluctuations produced by variations in the thickness of the capillary zone.

In line with this, Moore (1939) demonstrated that the magnitude of the potential variations depends on the rate of temperature change. In Moore's experiments, which were related to upward flow from a shallow water-table, the temperature range was limited to less than 3°C, and the greatest effect was registered with $\Delta T/\Delta t = -1$°C/hour. This resulted in a drop of the water-table of 3 cm in sand and 12 cm in clay.

Effect of Salts

Stallman (1964) pointed out that if an open container A with a salty solution is connected by means of a capillary U-tube with an open container B with pure water (Fig. 13a), there will be a continuous vapour-phase flow from B to A, causing dilution

and an increase in temperature in container A. The explanation of this phenomenon, which plays such an important role in the energy considerations of salt water conversion, is given by equation (8). In order to have completely static conditions in the unsaturated zone, the vapour phase must also be static. Referring to Figure 13:

$$H_A = -RT \log p_v/p_s$$
$$H_B = -RT \log p_v/p_0.$$

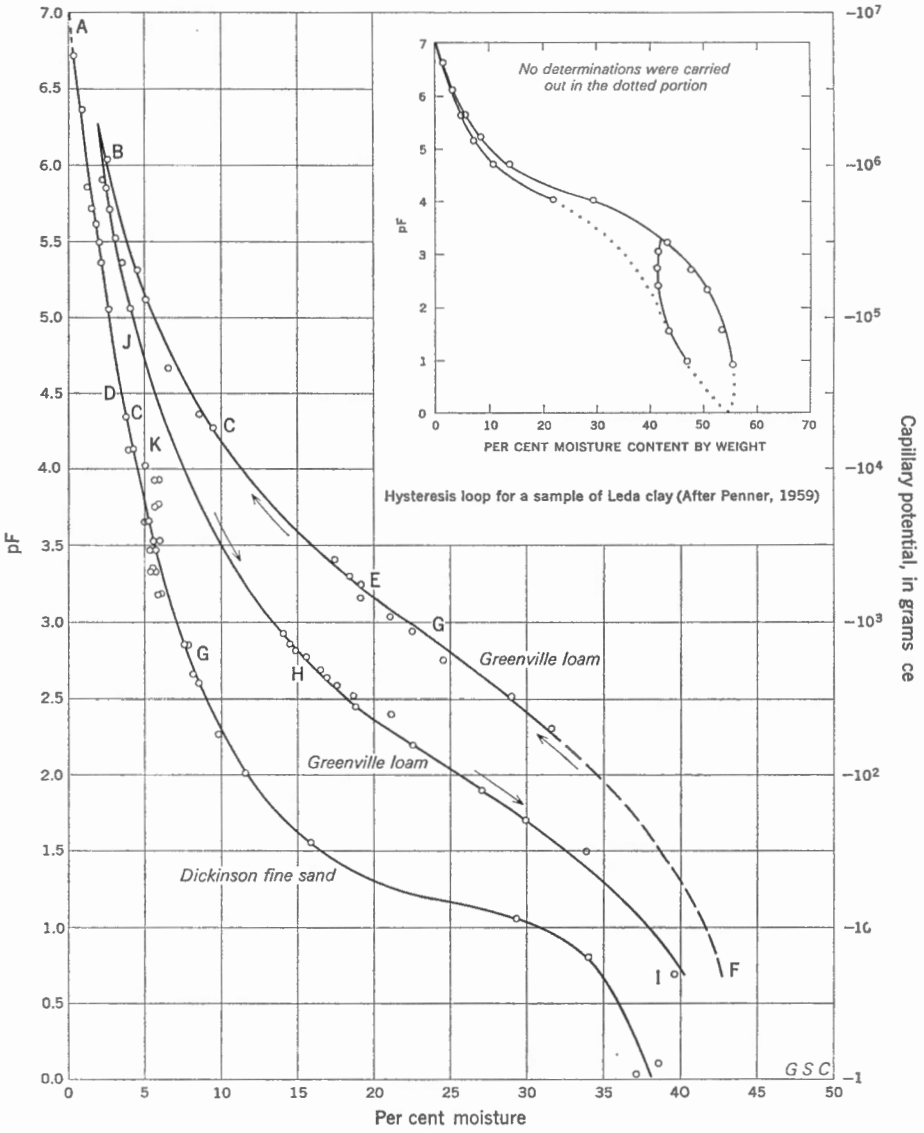


FIGURE 12. pH curves of some soils. The arrows indicate whether the soil is being progressively dried or moistened (*after* Baver, 1942).

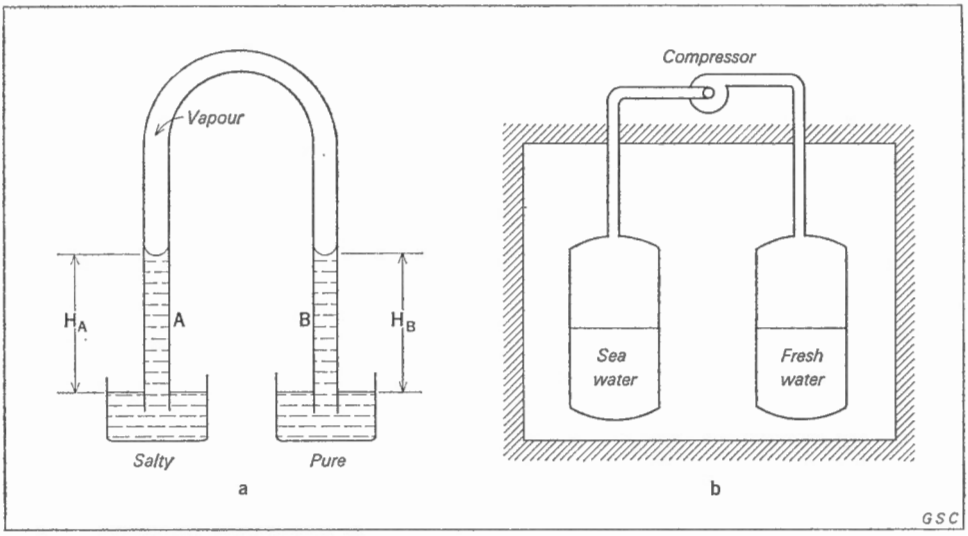


FIGURE 13. a, Vapour phase in vicinity of liquids (after Stallman, 1964); b, Schematic illustration of energy requirements for sea-water purification (after Spiegler, 1962).

The equilibrium vapour pressure of the salty solution (p_s) is less than that of pure water (p_0), which means that the vapour phase can not be static, and the observed flow results. Stallman mentioned that nonhomogeneous salt distribution in the unsaturated zone may play a significant role in the moisture distribution above the water-table.

Spiegler (1962, p. 15) calculated that although p_s is only 1.84 per cent less than p_0, the theoretical minimum amount of work necessary to transfer water vapour from the salt-water container to the fresh-water container is 0.70 kwh/m^3. This is also the theoretical minimum amount of work to desalt sea water or saline groundwater (Fig. 6b).

Empirical Coefficients Related to Certain Moisture Potentials

Comparison of agricultural literature, geological literature, and literature pertaining to petroleum engineering reveals that certain moisture contents have received much attention from all disciplines.

Veihmeyer and Hendrickson (1931) coined the term *field capacity* to indicate the lowest moisture content, expressed as percentage by weight, to which the soil may be brought by drainage alone, in a reasonable time (i.e., in a matter of days). The term *moisture equivalent* was introduced by Briggs and McLane (1907) to express the percentage of water, expressed in percentage dry weight, retained by a sample of soil when it is saturated and then subjected to a constant centrifugal force of 1,000 times that of gravity. Because the centrifugal force accelerates the effect that would have been produced by gravity drainage "in a reasonable time", its value may be somewhat larger than for field capacity. Nevertheless, its value may be taken as a reliable measure of the field capacity, particularly in fine-textured soils.

In general terms of moisture potential, moisture equivalent and field capacity are

characterized by a pF value of 2.7. Childs (1957) reports that field capacity can be attained with as much accuracy as the concept permits, at $\frac{1}{3}$ atm (pF 2.54) for light soils, increasing to $\frac{1}{2}$ atm (pF 2.71) for heavier soils.

A third empirical coefficient that is being used in agricultural studies is the *wilting coefficient*, which indicates the moisture content at which the vegetation supported by the soil starts to wilt. The wilting coefficient appears to be reproduceable with adequate precision at a moisture potential of 15 atm (Childs, 1957) or 4.19 pF. Peerlkamp (1949) mentioned the value of pF 4.15.

The range of soil moisture that is normally available to plant life lies within the range from pF 2.7 to pF 4.15. The actual amount of moisture that is stored in the soil between pF 2.7 and 4.15 within the 4-foot root zone depends on the soil texture and varies from 4.2 inches in sandy loam to 8.8 inches in heavy clay. This amount is generally referred to as normal *storage capacity*.

Field capacity is the upper limit of moisture available to plant life; conversely, it is the lower limit of moisture available for free drainage. For this purpose Meinzer (1923) introduced the term *specific retention*, which he defined as: "the volume of water which, after being saturated, the rock will retain against the pull of gravity to its own volume". It is stated as a percentage *by volume* and may be expressed by

$$R = 100\left(\frac{r}{v}\right) \qquad \text{(Meinzer, 1923) (9)}$$

where R=specific retention
r=volume of water retained
v=volume of rock.

To convert values of field capacity, expressed in per cent by weight, to specific retention, expressed as per cent by volume, it is necessary to multiply field capacity by the specific gravity of the rock or soil. Within the accuracy of the concept, specific retention can also be taken at a moisture potential of pF 2.7.

Figure 14 shows the relation between capillary pressure and water saturation of petroleum reservoir rocks. The relation is similar to the relation between moisture content and pF for soils, as is shown in Figure 12. Of interest to the petroleum engineer is the *irreducible water saturation* or *residual water*, which is the water content that can not be reduced by further increase in capillary pressure. Judging from Levorsen's graphs (Levorsen, 1954, Figs. 10, 11), the moisture potential corresponding to the irreducible water saturation is more than pF 4.

Measurement of Moisture Potential

In the laboratory, moisture potentials in the range pF 0–3 are generally determined by means of a suction plate apparatus (Penner, 1959). In the field, pF values of less than 2.8 can be determined by means of tensiometers such as those developed by Richards and his associates. A tensiometer consists of a porous cup filled with water and attached to a vacuum gauge. Instruments that register soil moisture contents of pF 2.7–4.3 on a volume per cent basis are based on the thermal properties of soil in moisture equilibrium with the soil water, or on the difference in neutron scattering of a dry and a wet soil.

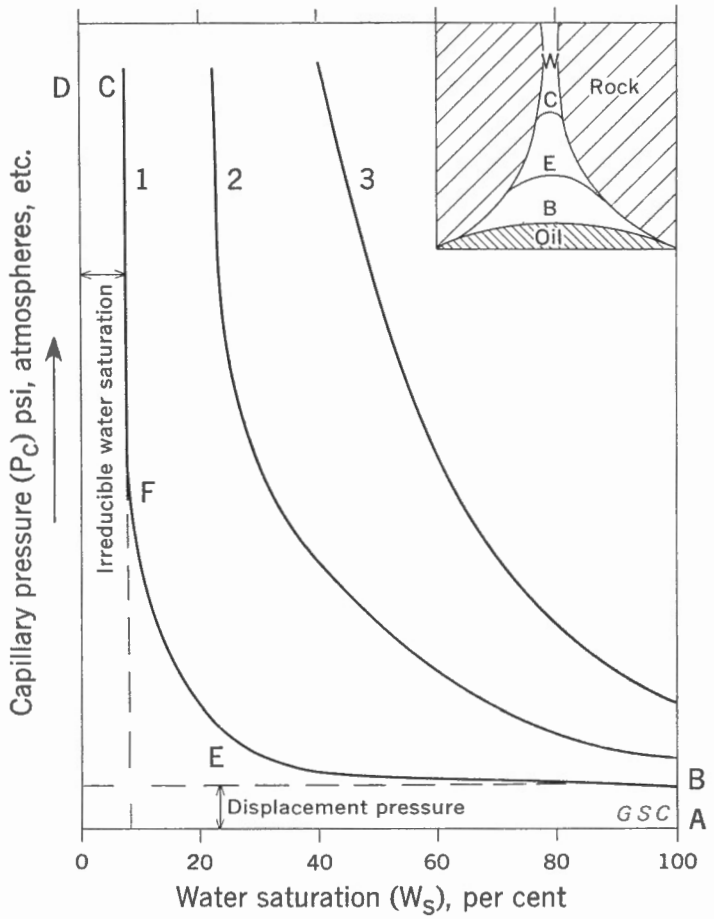


FIGURE 14.
Examples of capillary pressure curves for various reservoir rocks: 1, clean, well-sorted permeable rocks; 2, limestone or dolomite; 3, poorly sorted rock (*after* Levorsen, 1954).

Physical Characteristics of the Porous Medium and the Fluid

Permeability

Darcy determined experimentally that the amount of water that would flow through a vertical sand filter varies directly with the loss in hydraulic head (Fig. 15) according to

$$Q = KA\frac{H_1 - H_2}{L} \tag{10}$$

in which Q=amount of discharge

$\frac{H_1-H_2}{L}$=loss in hydraulic head over the length of flow L

A=cross-sectional area of the sand filter

K=constant of proportionality, depending on the permeability of the sand.

After Hagen (1839) and Poiseuille (1846), Hubbert (1940, p. 816) stated that the coefficient K expresses the relation of four different components as follows:

$$K=\frac{\rho g}{\eta}k' \qquad \text{(Hubbert, 1940, eq. 75) (11)}$$

where ρ=density of the fluid
g=acceleration due to gravity
η=viscosity of the fluid
k′=a property of the medium alone, pertaining to its ability to transmit water, and defined by Hubbert as the *coefficient of permeability*.

Recently, a committee of the Soil Science Society of America, under Dr. L. A. Richards, reviewed the terminology on permeability and presented the following recommendations (Luthin, 1957, p. 400):

1. Define *hydraulic conductivity* as the physical property which can be measured and expressed in terms of the Darcy K; that is, the proportionality factor in the equation $Q=KA\frac{H_1-H_2}{L}$.
2. If it is the author's preference to use the term permeability in connection with the Darcy K, make it clear at least once in the paper that hydraulic conductivity is the implied quantity.
3. Define permeability as the quantity measured and expressed by k′ in the equation $K=\frac{\rho g}{\eta}k'$ and, where necessary, use the compound term *intrinsic permeability* to avoid ambiguity.

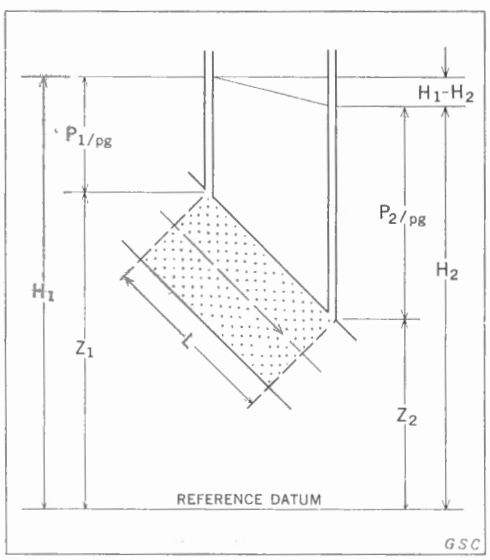


FIGURE 15.
Distribution of total hydraulic head in a uniform flow through porous medium.

In groundwater studies, the coefficient of permeability, denoted by P, is equivalent to hydraulic conductivity in agricultural usage. The term hydraulic conductivity deserves preference, not only because it avoids ambiguity, but also because of its analogy to the proportionality factors "electrical conductivity" and "thermal conductivity" in the corresponding electrical and heat flow equations.

The unit of permeability (intrinsic permeability) is defined by the American Petroleum Institute 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." This unit is the Darcy. (Levorsen, 1954, p. 101.)

Permeability as a Function of the Moisture Content; Relative Permeability

In the unsaturated zone, hydraulic conductivity is a function of the moisture potential. Hydraulic conductivity is very sharply reduced in the first stages of reduction of moisture content (Fig. 16). The effect is brought about by the entry of air into some of the pores, thus blocking them for moisture flow. The moisture content at which hydraulic conductivity reaches zero marks the beginning of the pendular stage. For this point, Moore (1939) gives the following pF values (Fig. 16): sand, pF 1.60; fine sandy loam, pF 1.96; silt, pF 2.08; clay, pF 2.16. Although other investigators report values up to pF 3, for most practical purposes zero hydraulic conductivity may certainly be taken at pF 2.7.

Moore's work led him to the formulation that, generally speaking, saturated permeability increases with increasing coarseness of texture in the following way:

sand > silt > clay.

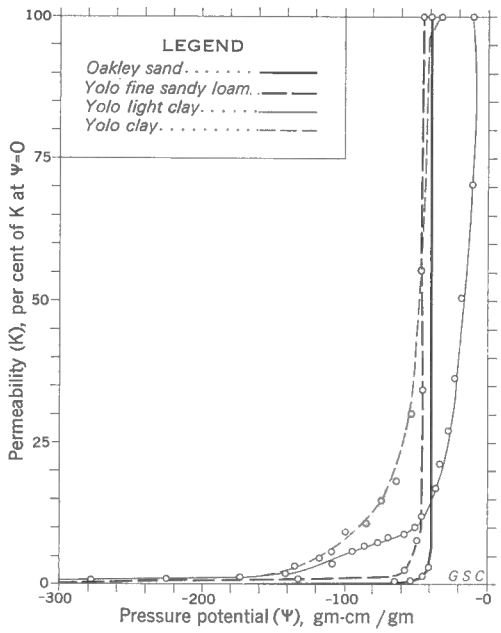


FIGURE 16.
Permeability as a function of moisture potential, with permeability expressed as a percentage of the permeability at a moisture potential of zero (*after* Moore, 1939).

In the range of unsaturated flow, the rate of change of permeability with moisture content increases with increasing coarseness in texture; in other words, unsaturated permeability decreases with increasing coarseness in texture as follows:

$$\text{sand} < \text{silt} < \text{clay}.$$

The ratio between effective permeability to a given fluid at partial saturation and the permeability at 100 per cent saturation is known in petroleum engineering as *relative permeability*. It is defined by Levorsen (1954, p. 107) as "the ratio of the amount of a specific fluid that will flow at a given saturation in the presence of other fluids, to the amount that would flow at a saturation of 100%."

In unsaturated and saturated permeability in soils, we were dealing with the presence of a multiphase fluid, consisting of a wetting liquid phase and a non-wetting gas phase. Schematic curves of relative conductivity for the liquid (saturated) and gas (unsaturated) phases are shown in Figure 17. Figure 18 shows a similar graph for relative permeability relations with varying saturations of oil and water.

The concept of relative permeability has wide application in the study of multifluid systems of petroleum, gas, and water. For instance, referring to the chart shown in

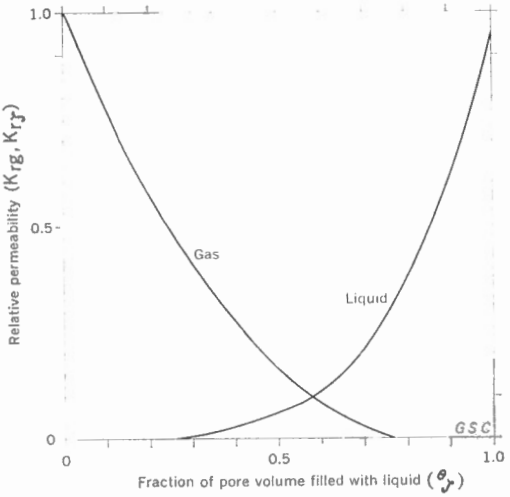


FIGURE 17.
Schematic relative conductivities for wetting liquids and gas phase (*after* Stallman, 1964).

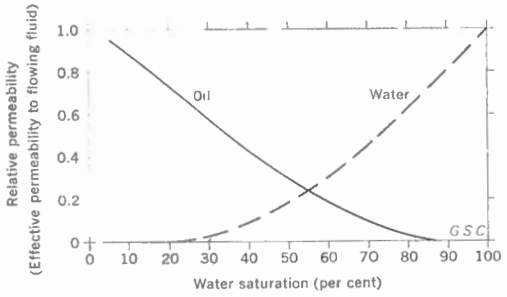


FIGURE 18.
Typical relative permeabilities with varying saturations of water and oil (*after* Levorsen, 1954).

Figure 18, Levorsen (1954) commented that for oil there must be a minimum of 5 to 10 per cent saturation of the pore space, before the non-wetting fluid (oil) can begin to move through the water-saturated rock and accumulate in pools.

Measurement of K

Hydraulic conductivity can be determined both in the laboratory and in the field. Standard laboratory measurements are done with a constant-head or falling-head permeameter, such as those described in most textbooks. Field methods are based on the rate of fluid rise in a drill-hole. A review of methods that are used in groundwater studies is given by Wenzel (1942), whereas the proceedings of the symposium on permeability in soils give a comprehensive review of methods used in agriculture and soil mechanics (A.S.T.M., 1954).

Isotropy and Anisotropy; Homogeneity and Heterogeneity

If the hydraulic conductivity at any point in a porous medium is the same in every direction, the medium is said to be *isotropic*. If the porous medium is isotropic and has the same hydraulic conductivity at all places, the medium is said to be *homogeneous-isotropic*. If the hydraulic conductivity at any point in the medium is not the same in every direction, and if this variation in hydraulic conductivity with direction is the same over the whole medium, the soil is said to be *homogeneous-anisotropic*. Thus, in a homogeneous anisotropic medium, the hydraulic conductivity is a function of direction but not of coordinates (Maasland, 1957, p. 216). If the hydraulic conductivity is different at different places in the medium, but if at any particular place the hydraulic conductivity is the same in all directions, the medium is said to be *isotropic-heterogeneous*.

In a qualitative description of a succession of horizontal laminae of thickness d, and hydraulic conductivities k_1, alternating with k_2, Childs showed the horizontal conductivity k_H to be

$$k_H = \frac{(k_1 + k_2)}{2} \tag{12}$$

and the vertical conductivity

$$k_V = \frac{2k_1 k2}{(k_1 + k_2)} \tag{13}$$

where the ratio k_H/k_V is called the *anisotropy quotient*.

Childs stated that "anisotropy is due to structural features which occur on a scale much smaller than that of the boundaries of the flow problem with which we happen to be dealing. If the thickness d is so small in comparison with the linear dimensions of the body of material as to be unnoticeable, then the material is to be regarded as anisotropic; but if the laminae were strata of thickness, say 50 cm, the problem would have to be treated as one of stratified isotropic soils, with conductivities k_1 and k_2 respectively."

It may also be expected that anisotropy exists in fissured rocks where one set of fissures is more abundant than the other and water may be conducted in their particular direction more freely than in other directions. Although Marinus Maasland (1957) has dealt extensively with the subject, anisotropy of earth materials has received little attention in theoretical groundwater studies. Hubbert (1940, p. 843) simply states, "we have agreed to consider only isotropic media."

Maasland advances the argument that "in a general way, it is more likely than not that a homogeneous natural porous bed is anisotropic as regards hydraulic conductivity and that, in general, one will find that $k_H > k_V$." As possible reasons for the natural anisotropy, Maasland listed: particle orientation, microstratification, or diagenetic changes in the sediment.

A porous medium will show isotropic-homogeneous properties with respect to fluid flow if the volume of the porous material is large compared to the pore sizes, so that the minute variations in intrinsic permeability, superimposed on the mean intrinsic permeability, are too small to be noticed. Of course, in dealing with the geological framework, homogeneity may never be expected to exist in a small area; nevertheless, at present, most hydrogeologists defend the thesis that, if taken on a large enough scale, rocks behave as a homogeneous isotropic medium with respect to fluid flow.

In his description of the hydrogeology of mountain ranges in northern Nevada, Farvolden (1964, p. 245) stated: "Homogeneity is a matter of scale. These mountains of highly deformed rock may well be just as homogeneous on a large scale as sand grains in a permeameter are on a small scale". Farvolden noticed that in order to meet the requirements of homogeneity, the minimum area considered should be two square miles.

Meyboom (1966) reasoned that glacial drift and Cretaceous shale in south-central Saskatchewan could be considered as homogeneous because:

> The distribution of pockets of sand and gravel in the till, and of the occasional sand lenses in the shale, is so completely at random, and the volumes of poorly permeable material are so large compared to the size of the gravel pockets and the sand lenses, that the variations of flow velocity, superimposed on the mean velocity are too small to be noticed, and it is a good approximation that both till and shale will show homogeneous-isotropic properties with respect to water flow.

Both reasonings are based on Hubbert's principle of the *macroscopic viewpoint*, according to which "we employ the macroscopic point of view if the fluid elements we speak of shall be large enough that the irregularities of flow due to the medium need not be considered, but only the statistical result" (Hubbert, 1940, p. 804; *see also* Hubbert, 1957, Fig. 5).

Fluid Flow Under Energy Gradients

Potential Energy Gradient

Reference has been made to Darcy's experiments, by which he showed that under full saturation the rate of water movement through a sand filter is proportional to the driving force and inversely proportional to the length of the path and the resistance to

flow exerted by the medium. In its generalized form, Darcy's law for three dimensional flow reads as follows:

$$V=-K\frac{dH}{dL} \tag{14}$$

where V=effective or macroscopic flow velocity[1]
K=hydraulic conductivity, as previously defined
$\frac{dH}{dL}=i$=hydraulic gradient.

The negative sign indicates that the flow is in the direction of decreasing hydraulic head.

The steady flow of fluid through porous media must obey, in addition to Darcy's law, the law of conservation of matter, which states that the volume of water that leaves a small saturated volume of the medium must equal the volume of water that enters the small volume during the same time. This so-called condition of continuity for steady state flow can be expressed mathematically as follows:

$$\frac{\partial V_x}{\partial x}+\frac{\partial V_y}{\partial y}+\frac{\partial V_z}{\partial z}=0 \qquad \text{(Peterson, 1957, eq. IV.7)}$$

in which V_x, V_y, and V_z are components of macroscopic velocity in the directions x, y, and z. The steady state condition implies that neither the boundaries of the flow system, nor the potentials along these boundaries, change with time.

The combination of Darcy's law with the continuity equation for steady state conditions results in the Laplace equation, which describes the spatial distribution of total hydraulic head (H) in an isotropic-homogeneous medium as follows:

$$\frac{\partial^2 H}{\partial x^2}+\frac{\partial^2 H}{\partial y^2}+\frac{\partial^2 H}{\partial z^2}=0 \tag{16}$$

By confining our attention to a two-dimensional flow region, the Laplace equation becomes:

$$\frac{\partial^2 H}{\partial x^2}+\frac{\partial^2 H}{\partial y^2}=0 \tag{17}$$

The distribution of total hydraulic head that is described by this equation is shown graphically in Figure 19, which is Hubbert's well-known flow pattern.

The Development of Flow Systems

Hubbert (1940, p. 843) described the flow pattern shown in Figure 19 as consisting of: "flow lines everywhere parallel to $-\text{grad}\,\Phi$ which form an orthogonal system with the family of equipotential surfaces: Φ=constant". Since we are limiting our attention to a two-dimensional plane through the flow field, the flow pattern shows a set of flow lines everywhere perpendicular to the equipotential lines in that two-dimensional plane. Along a single flow line, water will enter the region at a point of high potential (source) and it will leave at a point of low potential (sink). Along all stream lines, the

[1]The macroscopic flow velocity, which is the amount of discharge divided by the cross-sectional area through which flow occurs, is not to be confused with the actual flow velocity within the pores. If the porosity is f%, the actual flow velocity is equal to V/f.

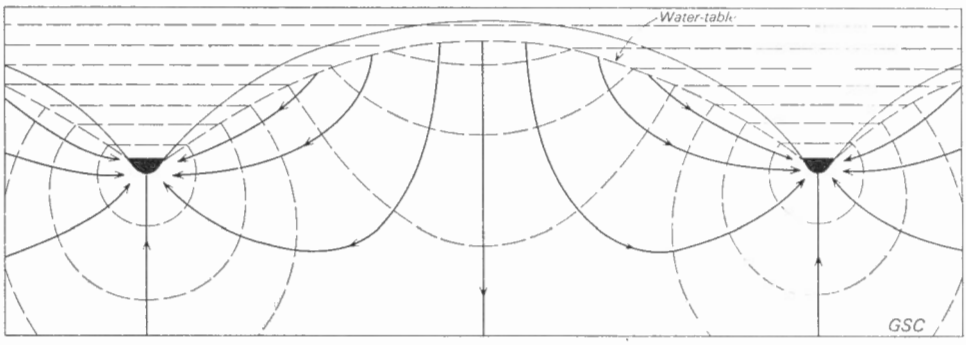


FIGURE 19. Approximate flow pattern in uniformly permeable material between the sources distributed over the air–water interface and the valley sinks (*after* Hubbert, 1940).

amount of work required to drive a unit mass of the fluid from one equipotential surface to another is precisely the same.

The combination of a homogeneous isotropic medium with a simple topographic configuration of a high area and an adjacent low area has been used widely as a model for groundwater flow in western Canada (Meneley, 1963). From this model, it was believed that groundwater discharge in the river valleys was a reliable estimate of all groundwater flow in a particular drainage basin implying that groundwater divides coincided everywhere with surface-water divides. Field evidence related to decreasing hydraulic head with increasing depth underneath topographic highs did indicate downward flow underneath the highs, but incongruities in the relation between total hydraulic head and depth below surface in the topographic low areas (Tóth, 1962) suggested that Hubbert's model was incomplete.

A more complicated flow model for groundwater flow in a small drainage basin[1] was calculated by Tóth (1962). His equation describing the distribution of fluid potential pertains to an area in which the following conditions should be fulfilled:

1. The basin is underlain by an impermeable boundary;
2. The two valley flanks are parallel and symmetrical relative to the stream;
3. The water-table is a subdued replica of the topography;
4. The geology is homogeneous and isotropic in the basin. The equation is also based on Laplace's equation, but by superposition of a sinusoidal surface on the regional slope of the basin, Tóth's analysis accounted for local topographic relief and the resulting potential distribution differs significantly from Hubbert's model. Under the given conditions, Tóth's potential distribution yields a local flow system, an intermediate flow system, and a regional flow system (Fig. 20), in which Tóth defined flow system as:

> A flow system is a set of flow lines in which any two flow lines adjacent at one point of the flow region remain adjacent throughout the whole region, and that can be intersected anywhere by an uninterrupted surface across which flow takes place in one direction only.

[1]Tóth defined a small drainage basin as an area bounded by topographic highs, its lowest part being occupied by an impounded body of surface water or by the outlet of a relatively low-order stream and having similar physiographic conditions over the whole of its surface.

Another important difference between Hubbert's model and Tóth's model is that, in the latter, groundwater discharge is not limited to the river valley but occurs between the "midline" and the valley bottom, in which midline is defined as "the line located on the surface midway between and parallel to the valley bottom and the water divide" (Tóth, 1962, p. 83).

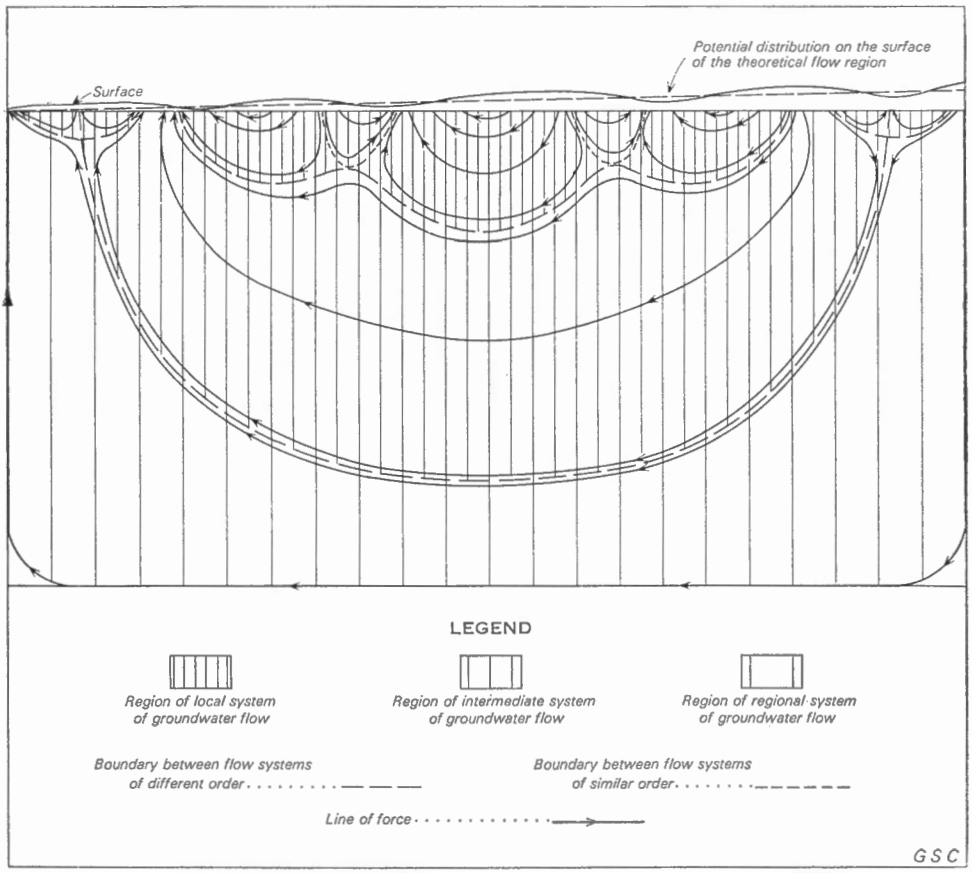


FIGURE 20. Theoretical flow pattern in a small drainage basin underlain by an impermeable boundary (after Tóth, 1962).

Based on a heatflow analogy, Tóth (1962) also calculated the distorting effect of extensive lenses of high permeability in a homogeneous medium of lesser permeability. From these calculations, it appeared that even small permeability ratios ($k'/k=10$) may have a pronounced effect on the flow pattern in the poorly permeable medium.

Meyboom (1962) measured the groundwater flow in a stratified medium consisting of a poorly permeable layer overlying a permeable layer. The flow pattern (Fig. 21) showed predominantly downward movement through the poorly permeable layer, lateral movement in the permeable layer, and radial movement in the vicinity of a

drainage channel. Because of its general occurrence in western Canada, the model was called the Prairie Profile, which Meyboom described as follows:

> The Prairie Profile consists of a central topographic high, bounded at either side by an area of lower elevation. Geologically, the profile is made up of two layers of different permeability, the upper layer having the lower permeability. Through the profile is a steady flow of groundwater from the area of recharge to the area of discharge. The ratio of permeabilities is such that groundwater flow is essentially downward through the material of low permeability and lateral and upward through the underlying more permeable layer. The potential distribution in the profile is governed by the differential equation of Laplace.

The fact that the direction of flow in the poorly permeable layer tends to be more vertical than that in the permeable layer is explained by Casagrande's law of refraction (Casagrande, 1937) of stream lines at a boundary between strata of different hydraulic conductivities (Fig. 22). This law of refraction can be expressed as:

$$K_1 \cot \theta = K_2 \cot \psi$$

where K_1 and K_2 are hydraulic conductivities and $K_1 > K_2$
θ=angle between the flow line in the permeable material and the vertical
ψ=angle between the vertical and the flow line in the poorly permeable material.

Hubbert (1953, p. 1971) formulated Casagrande's refraction law as follows:

> In an alternating series of sand and silt strata, any flow oblique to the bedding would occur in sharply zig-zag paths, traversing the more permeable strata nearly parallel with the bedding, and the less permeable strata more nearly at right angles.

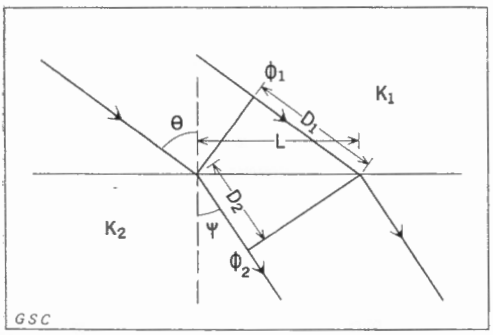


FIGURE 22.
The refraction of stream lines at the boundary between two media of different hydraulic conductivities K_1 and K_2 (*after* Childs, 1957).

The patterns of groundwater flow in the Prairie Profile show a striking resemblance to theoretical flow patterns calculated by Ernst (1962) for drainage through stratified soils. Subsequent field observations in areas where the Prairie Profile was expected to be a valid model indicated that groundwater discharge is indeed limited to the drainage channel, provided that "the flanks of the river valleys are 5 to 15 times steeper than the regional slope between the watershed and the river" (Meyboom, *et al.*, 1966).

The development and understanding of groundwater flow systems have added significantly to the usefulness of regional groundwater studies, for the proper understanding of the flow system is prerequisite both to intelligent measurement of the groundwater balance and to simulating the groundwater flow in a model.

Unsaturated Flow along a Potential Gradient

Laboratory investigations by Hansen (1955) pertaining to unidirectional water movement along a potential gradient through unsaturated soil indicated the existence of three zones that occur as a dynamic continuum. Hansen called these zones *transmission zone*, *wetting zone*, and *wetting front*.

Within the so-called transmission zone, the moisture content is essentially constant, regardless of the direction of flow. The degree of saturation is approximately 80 per cent. The constant moisture content assures a constant hydraulic conductivity.

The moisture content in the wetting zone reduces rapidly as the wetting front is approached, causing a corresponding decrease in hydraulic conductivity. Hansen commented that "visual observation of the wetting phenomena which take place within the wetting zone indicates that flow occurs in a very erratic manner, the movement taking place essentially in increments which might be described as 'jumping' movement." From the experiments, the wetting front appeared to be an effective restrictive layer to downward movement of water.

From the observation that the pressure potential throughout the transmission zone during downward flow is *constant*, Hansen derived the relation that for downward flow the rate of infiltration approaches a constant as time increases (Hansen, 1955, eq. 23), for upward flow the rate of flow approaches zero with increasing time. It should be emphasized that the constant pressure head during the infiltration indicated that the decrease in total hydraulic head in the transmission zone is brought about entirely by the change in gravitational energy (elevation head).

Other moisture flow in unsaturated media is either along a temperature gradient or along an osmotic gradient, both of which are discussed briefly hereafter.

Temperature Gradient

Thermal transfer of moisture in unsaturated porous media is of importance at a very narrow range of liquid contents and is always in the form of vapour transmission. First experiments by Bouyoucos (as quoted by Smith, 1943) led to the conclusion that thermal moisture transfer was maximal at a certain moisture content, which Bouyoucos called "thermal critical moisture content." Smith (1943) improved the measuring technique of Bouyoucos and substantiated Bouyoucos's conclusions at much lower temperature gradients (1.8°–5.5°C/cm as compared to 20°C/cm in Bouyoucos's experiments).

Smith (1943, p. 520) stated,

> The most perplexing observation is that the thermal transfer of moisture begins to be appreciable at a moisture content just above that required for equilibrium, with a relative humidity of 75%, and then practically ceases at about moisture equivalent. . . . The whole effect appears to be confined to the region of funicular water.

Expressed in terms of moisture content in loam (Fig. 23), thermal moisture transfer is negligible at 5%, rises to a maximum at 14%, and is again nearly zero at 25%. Smith concluded from this that the vapour transfer in the funicular zone induces a capillary gradient in the capillary zone, resulting in phreatic water changing into capillary water. The flow increases with increasing temperature gradient.

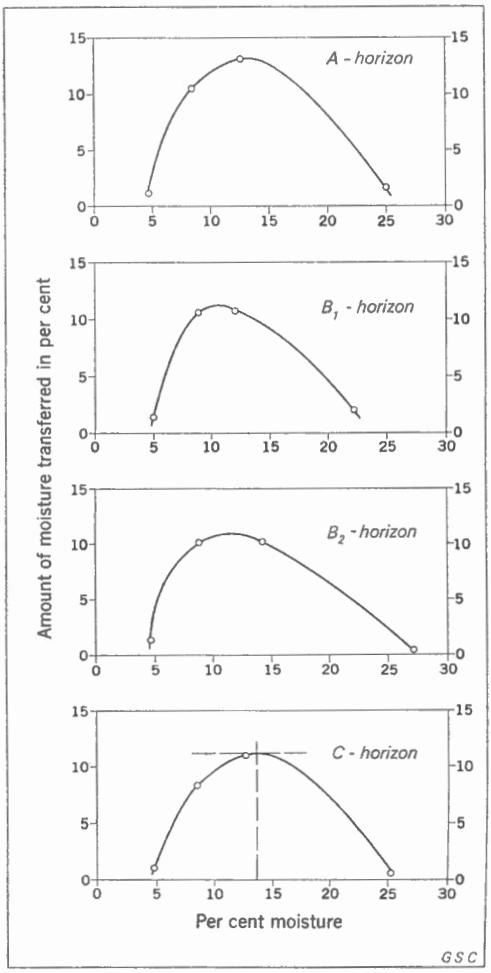


FIGURE 23.
Amount of moisture transfer as a function of moisture content (*after* Smith, 1943).

The importance of this phenomenon under field conditions was demonstrated by Schneider (1961), who correlated air temperatures with water level fluctuations. From these measurements, he concluded that during the winter thermal moisture transfer is the chief mechanism by which groundwater is being transported upward towards a growing frost wedge, resulting in accretion to the frost layer from below (Fig. 24). The total upward moisture movement during the winter may cause a water-table decline of up to 9 feet.

Osmotic Gradient

Movement of moisture along an osmotic gradient is of importance in the range of moisture potentials between pF 2.7 and pF 4.2, where we are dealing with the influence of plant life.

Evaporation from the plants causes an increase in cell-sap concentration in the cells surrounding the stomata. The increasing concentration induces an osmotic

gradient between adjacent cells, and innumerable compensating currents are set up, combining in a single ascending water flow as long as evaporation continues. So, the principal role of the osmotic gradient is in the transfer of moisture from the soil into the plant through the cell membranes. Although the process is poorly understood analytically, several investigators have established the existence of an osmotic gradient between plant and soil and within the plant, as well as the increase in the overall osmotic value of the cell sap during dry periods.

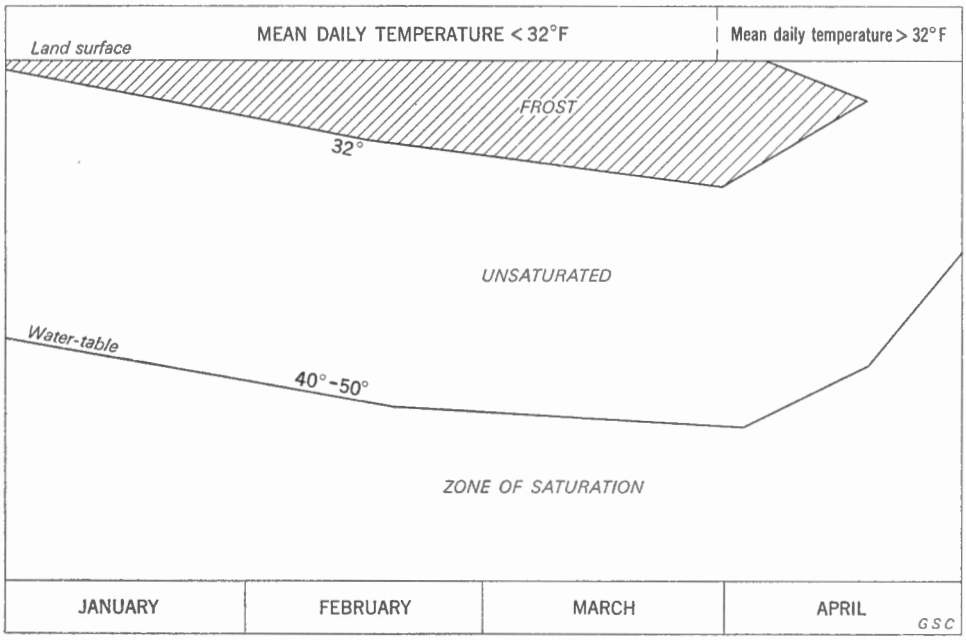


FIGURE 24. Sketch illustrating thermal conditions, the trend of the water-table, and the occurrence of frost in the winter and spring (*after* Schneider, 1961).

The Principle of Continuous Moisture Flow

Groundwater flow is part of the hydrological cycle. It is possible to establish the water balance for each part of this cycle according to the following general equation:

$$Q = kf\left(\frac{dH}{dL}\right)F \qquad (19)$$

where Q=constant quantity of flow
F=cross-sectional area through which water flows
dH=difference in total hydraulic head over a short length dL of the flow path
f=arbitrary function
k=hydraulic conductivity.

Flow through the hydrologic cycle involves phase transitions, and it is therefore necessary to combine Darcy's law and the continuity equation with the appropriate physical and chemical reaction laws that govern the individual phases. By doing so, it

is possible to describe the entire hydrologic cycle in terms of a *continuous moisture flow* (Visser and Bloemen, 1960). The approach taken by Visser and Bloemen is highly imaginative, but difficult to measure. By elaborating on equation 19, they state:

The value of Q for a certain section is the same for the next section if no changes in storage take place. The relation may be rewritten as:

$$\frac{f_1(dL_1)}{F_1}Q=k_1f_1(dH_1)=k_2f_2(dH_2)\frac{f_1(dL_1)F_2}{f_2(dL_2)F_1},$$

in which they assume that $f_1\left(\frac{dH}{dL}\right)=\frac{f_1(dH_1)}{f_1(dL_1)}$, which for an arbitrary function is not necessarily true.

Nevertheless, the "moisture flow technique" of Visser and Bloemen may lead to a more fundamental approach in all hydrology studies. Although speculative, at present it is the only attempt to unify the entire hydrologic cycle into one mathematical expression.

Part 2: Methods in Hydrogeology

Hydrogeology, as a science, deals with nature, and "nature" means "the actual world". The concern of a science about its methodology is directly proportional to the concern of that science with nature, as is demonstrated by the following sequence: logic → mathematics → physics → chemistry → (hydro) geology → biology → medicine.

The method of natural sciences in general is materialistic in that it is based on the hypothesis that all processes and phenomena are interrelated. In contrast to the metaphysical view, according to which nature is a fortuitous accumulation of objects, the materialistic viewpoint holds that nature is a connected single whole, where objects and phenomena are organically bound to one another, depend on one another, and cause one another.

The methods that are used in hydrogeology, as in other natural sciences, to unravel this complex of interrelations depend on observation and classification, followed by two complementing procedures of thought: induction and deduction.

The Cycle of Research

The Principle and a Simplified Example

Reviewing the papers that were presented at a congress dealing with the method in science, Schouten (1960) summarized this method as follows: Through a systematic arrangement of observations, we arrive at an empirical rule; this is followed by induction and abstraction, which lead to the formulation of a hypothesis, theory, or postulate. By deducing certain consequences from the theory, it is possible to predict the outcome of renewed observations, which in turn can be applied to judge the probability of the theory in relation to the evidence or to review the original empirical rule. Schouten called this never-ending process of examination and re-examination the "cycle of research" (Fig. 25). In geology, van Bemmelen (1961) called this the "prognosis-diagnosis method", consisting of an "ascend" from the isolated facts to the working hypothesis, followed by a "descend" from the concept to the facts.

The cycle of research can best be illustrated by an example. Figure 26 shows the cycle leading from Darcy's law to the present knowledge of flow systems. Darcy's empirical rule resulted from a very practical problem related to the design of a suitable sand filter for the public water works of the city of Dyon. By induction and further experimenting, Pennink (1905) determined from Darcy's law the fundamental pattern of groundwater flow in the vicinity of a drainage canal. Later, Pennink's flow pattern was substantiated theoretically by Hubbert (1940), who, after examination of the physical content of Darcy's law, developed the theory that groundwater moves in the direction of diminishing fluid potential and that the distribution of the fluid potential is determined by the Laplace equation.

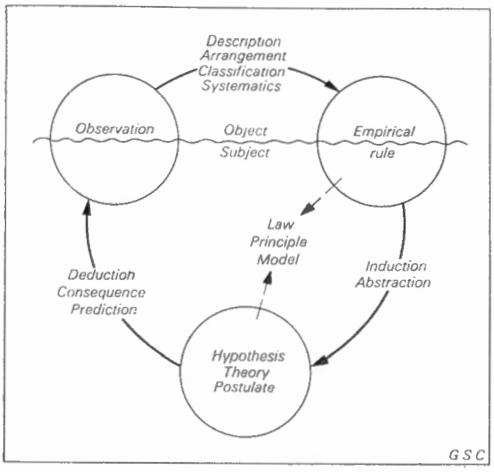


FIGURE 25.
The cycle of research (simplified after Schouten, *in* Bottema, *et al.*, 1960).

Renewed observations, aided by laboratory experiments and simple potential measurements, led to the mathematical description of more complicated models of groundwater flow (Edelman, 1950; Ernst, 1962; Tóth, 1962). As a result of the new evidence, new empirical models emerged, suggesting that groundwater flow systems may exist on a continental scale (Meyboom, 1962).

These models, combined with the recognition of the hydrologic cycle in hydrogeology, have made the water balance (or water budget) the final phase of regional groundwater studies (*see* Table I).

This review of the origin of an approach is too brief to do justice to the many-sided evidence on which the individual elements of the idea are founded. The following summary will elucidate some of the links shown in the cycle of Figure 26.

Observation and Classification

First and foremost in hydrogeology is the knowledge of the geological framework. No judgment as to the applicability of physical principles can be passed without an understanding of the materials and their spatial arrangement. Geology is the prerequisite for all further study. However, geological knowledge has to be augmented by

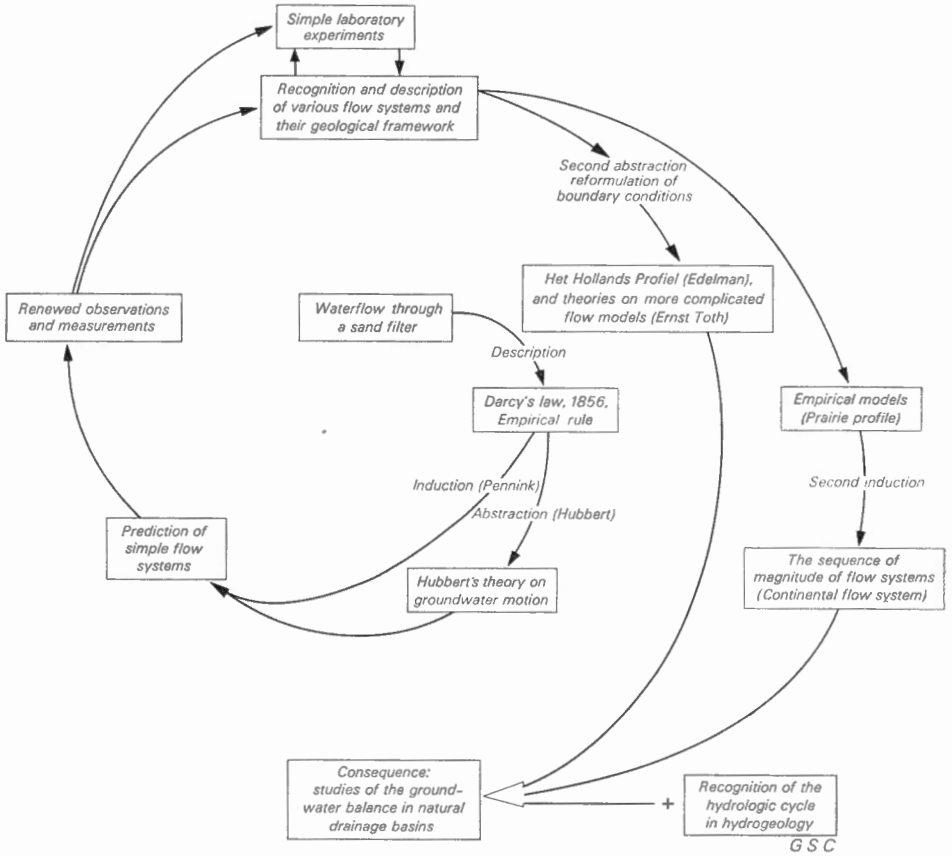


FIGURE 26. Simplified example of the cycle of research related to one aspect of hydrogeology.

direct and indirect observations pertaining to the hydrology of the underground water.

The most important direct observations are aquifer tests and piezometric cross-sections, whereas indirect observations may be chemical or geobotanical, based on the premise that all natural processes and phenomena are interrelated.

Aquifer Tests

The technique of aquifer tests, dictated as it is by immediate water supply problems, has advanced much faster than any other branch of hydrogeology. The Theis equation, the modified Theis equation, and the various leaky aquifer equations (Walton, 1962) are well-known tools in aquifer evaluation. However, this development has taken place in isolation from the rest of the science, particularly with respect to the natural water balance.

This lack of concern for the natural groundwater replenishment is reflected in the concept of "safe yield" of a water well; that is, the pumping rate at which the draw-

down for each logarithmic time cycle (1–10 minutes, 10–100 minutes, etc.) does not exceed one eighth of the total available drawdown. Walton (1964) reports that such pumping rates have caused a drop in the water levels of deep wells in Illinois of 650 feet since 1864, and he warns that safe physical yield is not identical with *safe sustained yield.*

Piezometer Cross-sections

Piezometer cross-sections afford a means of obtaining direct values of total hydraulic head, from which flow patterns can be constructed (Meyboom, *et al.*, 1966). The technique was first used by Pennink (1905) and is now among the standard instrumentation of the Geological Survey of Canada. The resulting flow patterns enable a better assessment of the various discharge and recharge patterns in a certain area than conventional "piezometric maps".

Hydrochemistry

After determining and comparing a large number of analyses of natural waters, Chebotarev (1955) formulated the following empirical rule pertaining to the chemical composition of groundwater: "In general, in the area between the water divide and the erosion basis the following sequence can be distinguished: bicarbonate water (HCO_3^-), sulphate water ($SO_4^=$), chloride water (Cl^-)." The chemical explanation for this sequence is given in Chebotarev's sixth conclusion which reads:

> While the least soluble salts are precipitated first and the most soluble salts last, at any given time and at any distance from the intake area the chemical components of higher solubility will be found in water in greater relative abundance. The sodium chloride tends, therefore, to be in solution as long as the extremely high salinity concentration of water is achieved (35–42%).

Hydrodynamically, Chebotarev relates the predominant types of groundwater to the hydraulic gradient and distinguishes between (a) the zone of active exchange, (b) the zone of delayed exchange, and (c) the zone of stagnant conditions.

Schoeller (1959) lists the same sequence of predominant ions as follows:

(1) $rCO_3 > rCl > rSO_4$
up to 60 milliequivalents

(2) rCl or $rSO_4 > rCO_3$
up to 180 milliequivalents

(3) $rCl > rSO_4 > rCO_3$
and $rNa > rMg > rCa$.

Chemically, Schoeller (1959, pp. 54–85) recognizes dissolution in the upper level of the mother rock as the process most responsible for the composition of groundwater. Dissolution is followed by three modifying phenomena: reduction, base exchange, and concentration. Schoeller suggests a slow circulation as a possible hydrodynamic mechanism for these processes.

It will be obvious why these observations on hydrochemistry have aided greatly in the further understanding of flow systems. Chemically, the sequence in composition

reflects a sequence in time, and hence, since groundwater is moving, a sequence in place. In terms of a flow system, the chemical successions of Chebotarev and Schoeller may be expected to coincide closely with chemical changes of groundwater within a flow system. The degree to which the sequence of alteration has advanced depends on the magnitude of the flow system.

Another important family of hydrochemical observations are tracer studies, which use the stage of radioactive decay of natural or artificially introduced radioactive substances to trace the flow path of water. Substantial advances in such techniques have been published recently by the International Atomic Energy Agency in Vienna (1964).

Geobotany

Geobotany, as a method of classifying groundwater occurrences, was applied as early as 1900 (Mager, 1912). The principle of the method is well defined by Viktorov (1955) in his description of "hydroindicatory investigations". He stated that "any phenomenon in nature may be understood and substantiated if it is seen in its continuous connection to the surrounding phenomena, in its stipulated dependence on the phenomena surrounding it."

Meinzer (1923), Robinson (1958), Viktorov (1955) and Vostokova (1961) showed the applicability of botanical observations to groundwater studies. In the beginning, these studies, which related groundwater chemistry or groundwater proximity to vegetation, were simply means of classification. During the stage of renewed observations, when it became accepted to associate groundwater chemistry with stages in the flow system, geobotany could be applied as a means of recognizing surface manifestations of a flow system ("Mapping the flow system", Meyboom, 1962).

This development stresses the fact that renewed observations need not be new observations, but may simply be the re-examination of old observations in the light of a new hypothesis.

Laboratory Experiments

Experiments are never proof that a natural process took place analogously or will take place analogously. Nevertheless, they stimulate the imagination and afford a means to test simple ideas. Laboratory experiments in hydrogeology range from Darcy's sand filter to the present electrical analog models.

Two types of models can be distinguished: hydraulic scale models and analog models. In the hydraulic scale models nature is imitated as closely as possible on a reduced scale. Analog models are based on the principle that two apparently unrelated physical phenomena are analogous if their differential equations—each with its own boundary conditions specified in terms of independent variables—are formally identical. As Darcy's law is formally identical with Ohm's law and Poiseuille's law each one of which is expressed by the Laplace equation groundwater flow may be simulated by electrical current in a uniformly conductive medium or by the flow of a viscous fluid through a narrow channel. In a recent review, van Everdingen and Bhattacharyya (1963) list all techniques and principles that are being applied to groundwater model studies.

Inductive and Deductive Methods

Once the arrangement of measurements and observations has led to the formulation of an empirical rule, two ways of generalization lie open: induction and abstraction. What may be expected from either way of thinking?

Induction is the apprehension of the universal in the particular (Eaton, 1959), and it leads to the construction of systems of generalizations from particular cases, refined and supported by both positive and negative evidence. Eaton called this the "primary phase of induction." The secondary phase of induction starts when generalizations reinforce one another by entering logically organized systems, the value of which is confirmed by more or less remote consequences deduced from them.

Inductive generalizations can never be proved. Proof, according to Eaton, is the process by which some proposition is made credible through its logical relation to other propositions; i.e., becomes a premise. Although the premises of an induction are true (being drawn from actual fact), it is never certain whether its generalizations are true. The more precise the observations are, however, the more safely the generalizations can be made.

Abstraction translates observation into certain idealized ("theoretical") assumptions for a deductive system which lays down certain rules to which particulars must conform if they belong to that system. However, the system does not deduce the *existence* of the particulars. As the premises of a deductive system are not arrived at by any process that makes them credible through their relation with other propositions, the deductions are as true as the premises, and it is not sufficient to *assume* the truth of the premises.

Eaton (1959, p. 484) summarizes the two ways of thinking as follows:

> Inductive and deductive procedures are the inverse of one another in this sense: the generalizations of a deductive system prescribe conditions to which particulars must conform if they are in that system; the generalizations of an inductive system are prescribed by the particulars antecedently assumed to be in that system.

It is shown clearly in Figure 26 how either way of thinking is employed in the development of a science.

The Sequence of Activities

Table IV shows the succession of activities leading to a complete understanding of groundwater occurrences. The five phases shown in this table correspond with the succession of thoughts that led to the present concept of flow systems, and they can be described briefly as follows:

1. The first phase, comprising the determination of the geological framework, falls within the realm of classical geology. It consists of surface and subsurface investigations pertaining to the stratigraphy and structure of an area. It also includes determinations of the physical properties of the materials (permeability and porosity) by means of field or laboratory tests. The products of the first phase are groundwater probability maps.

Table IV

Sequence of Activities Leading to Complete Understanding of Groundwater Resources

Phase	*Place and Nature of Activity*	*Details*	*Origin of Information*	*Relevancy*	*Products*
1. Geological framework	Field (Mapping, drilling, geophysical methods) *Regardless of physical boundaries*	Stratigraphy Structure Determinations of physical properties: *pump testing*	Geological Survey of Canada (may or may not be within groundwater section) Provincial Geological Surveys Research Councils, etc.	—Phase 1: Probability study	Groundwater probability maps
2. Manifestations of the flow system	Field (piezometers, measurement of key wells, geochemical and geobotanical mapping of discharge areas)	Topography— Maximum elevation = maximum fluid potential = water divide Minimum elevation = minimum fluid potential = river This defines the boundaries of the basin		—Phases 1–3: Reservoir and Damsite investigations; studies related to the feasibility of irrigation pollution, stream diversion	Groundwater reports on basins, areas, and special studies
3. Construction of three-dimensional flow system	Office (laboratory)	Distribution of recharge and discharge areas Relation between extent of selected basin and depth of the flow system (Models)	*Phases 1 and 2*		Groundwater flow diagrams and maps
4. Quantitative determinations at the boundaries of the flow system	Field + Office processing of available outside data	Input: Precipitation (*Infiltration*); Artesian feeding Output: Run-off (surface and underground); Evaporation; Surface water groundwater; Transpiration	Geological Survey of Canada Inland Waters Branch Meteorology Branch D.O.T Dept. of Agriculture P. F.R.A. Provincial Water Agencies	—Phases 2–5: are popularly referred to as "*Basin Study*"	Preliminary water balance estimates
5. Water balance	Office	Equation of Input and Output: Distribution of availability	*Phases 1 to 4*		Groundwater availability maps

2. The second phase involves the qualitative determination of the flow system by means of piezometer measurements, aided by geochemical and geobotanical methods. The area of investigation is no longer undetermined but becomes confined within the boundaries of highest fluid potential, which in simple cases coincide with drainage basin boundaries.

3, 4. The third step is the construction of the flow system from all available information, followed by the fourth step, which involves measurements at the boundaries of the flow system, in order to determine the quantitative aspects of the flow.

5. The final phase contains the water balance (or water budget) which expresses the natural balance between groundwater recharge and groundwater discharge, and the resulting map may be called groundwater availability map.

Table IV shows that certain specialized studies are made up of only one or two of these steps, depending on the problem to be solved. From the literature references in this table, it can be seen to what extent the various phases of this scheme have been studied by the Geological Survey of Canada.

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Chapter III

APPALACHIAN HYDROGEOLOGICAL REGION

by P. A. Carr

Introduction

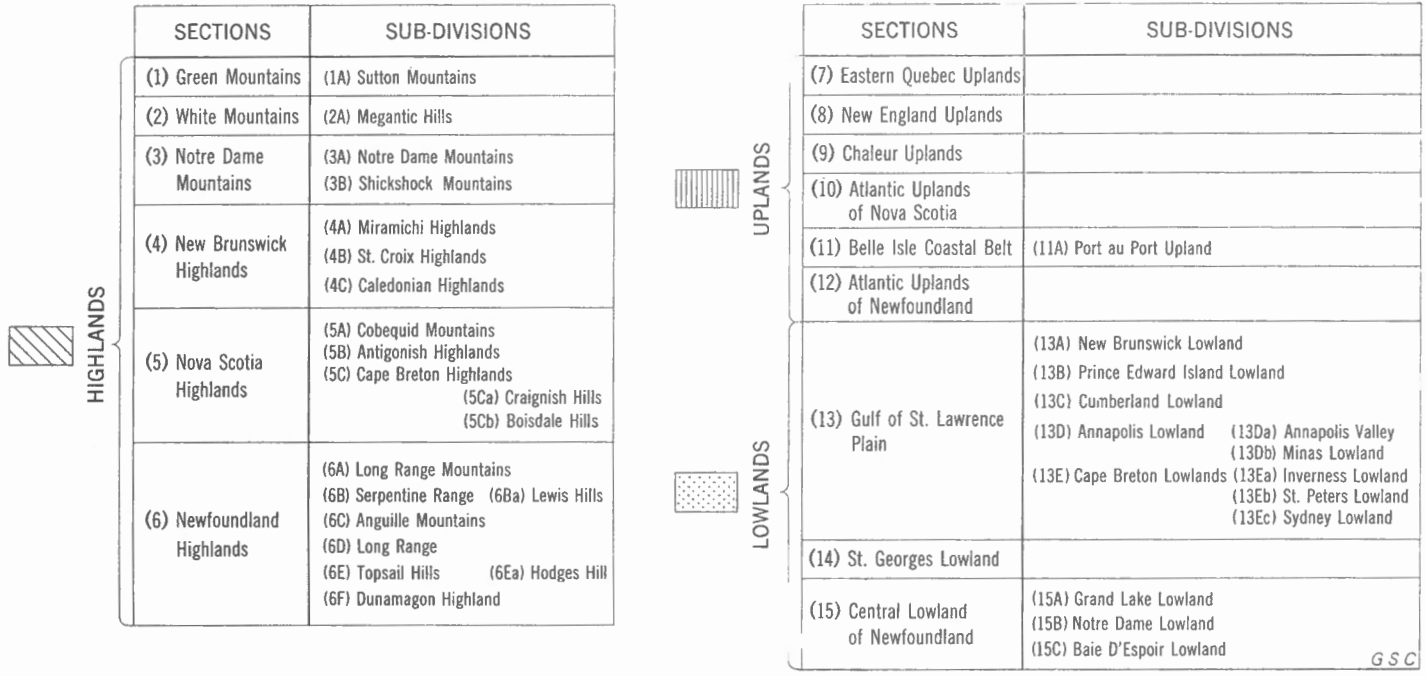
The purpose of this chapter is to describe relatively the quantity and quality of groundwater in the rocks of the Appalachian area. This is done by summarizing the published and the available unpublished reports on the subject.

The Appalachian hydrogeological region of Canada includes that part of Quebec that lies southeast of the Logan fault (Fig. 7), as well as the Magdalen Islands, all of New Brunswick, Prince Edward Island, Nova Scotia, and Newfoundland. This is part of a larger physiographic region, the Appalachian mountain system, which extends from Newfoundland to Alabama. The physiographic divisions of the Canadian Appalachians and their relative topographic position are shown in Figure 28.

The Appalachian hydrogeological region receives more rainfall than any other part of Canada except the western half of the Cordilleran region. In general, the quantity of groundwater available is not large, for the rocks are not very permeable. The bedrock comprises mainly igneous and metamorphic rocks of a pre-Carboniferous age and sediments of Carboniferous, Permian, and Triassic age (Fig. 29). The surficial deposits of sand and gravel are good aquifers in many places but often are small, and limited in areal extent.

Prior to 1930, groundwater was mentioned in the literature only in an occasional reference to good springs, saline springs, or spring lines at lithological contacts of differing permeability. One of the earliest references according to Brandon (1966) was that of Colonel Franquet (1751), who described several excellent springs in a report discussing the economy and fortifications of Prince Edward Island. Hind (1865), Chalmers (1890), Bailey (1898), and Ells (1907) described saline springs in various parts of New Brunswick, some of which they had analyzed (Hamilton, 1961). Hayes (1931), when prospecting for potash in Nova Scotia, studied the geology of saline springs and had them analyzed.

Reports on the groundwater resources of parts of Prince Edward Island were published by Owen and Pollitt (1950) and by Pollitt (1952 and 1953). These reports relatively assessed the bedrock and surficial aquifers and determined the chemical quality of the groundwater.



	SECTIONS	SUB-DIVISIONS
HIGHLANDS	(1) Green Mountains	(1A) Sutton Mountains
	(2) White Mountains	(2A) Megantic Hills
	(3) Notre Dame Mountains	(3A) Notre Dame Mountains (3B) Shickshock Mountains
	(4) New Brunswick Highlands	(4A) Miramichi Highlands (4B) St. Croix Highlands (4C) Caledonian Highlands
	(5) Nova Scotia Highlands	(5A) Cobequid Mountains (5B) Antigonish Highlands (5C) Cape Breton Highlands (5Ca) Craignish Hills (5Cb) Boisdale Hills
	(6) Newfoundland Highlands	(6A) Long Range Mountains (6B) Serpentine Range (6Ba) Lewis Hills (6C) Anguille Mountains (6D) Long Range (6E) Topsail Hills (6Ea) Hodges Hill (6F) Dunamagon Highland
UPLANDS	(7) Eastern Quebec Uplands	
	(8) New England Uplands	
	(9) Chaleur Uplands	
	(10) Atlantic Uplands of Nova Scotia	
	(11) Belle Isle Coastal Belt	(11A) Port au Port Upland
	(12) Atlantic Uplands of Newfoundland	
LOWLANDS	(13) Gulf of St. Lawrence Plain	(13A) New Brunswick Lowland (13B) Prince Edward Island Lowland (13C) Cumberland Lowland (13D) Annapolis Lowland (13Da) Annapolis Valley (13Db) Minas Lowland (13E) Cape Breton Lowlands (13Ea) Inverness Lowland (13Eb) St. Peters Lowland (13Ec) Sydney Lowland
	(14) St. Georges Lowland	
	(15) Central Lowland of Newfoundland	(15A) Grand Lake Lowland (15B) Notre Dame Lowland (15C) Baie D'Espoir Lowland

Key to Figure 28 (*facing page*)

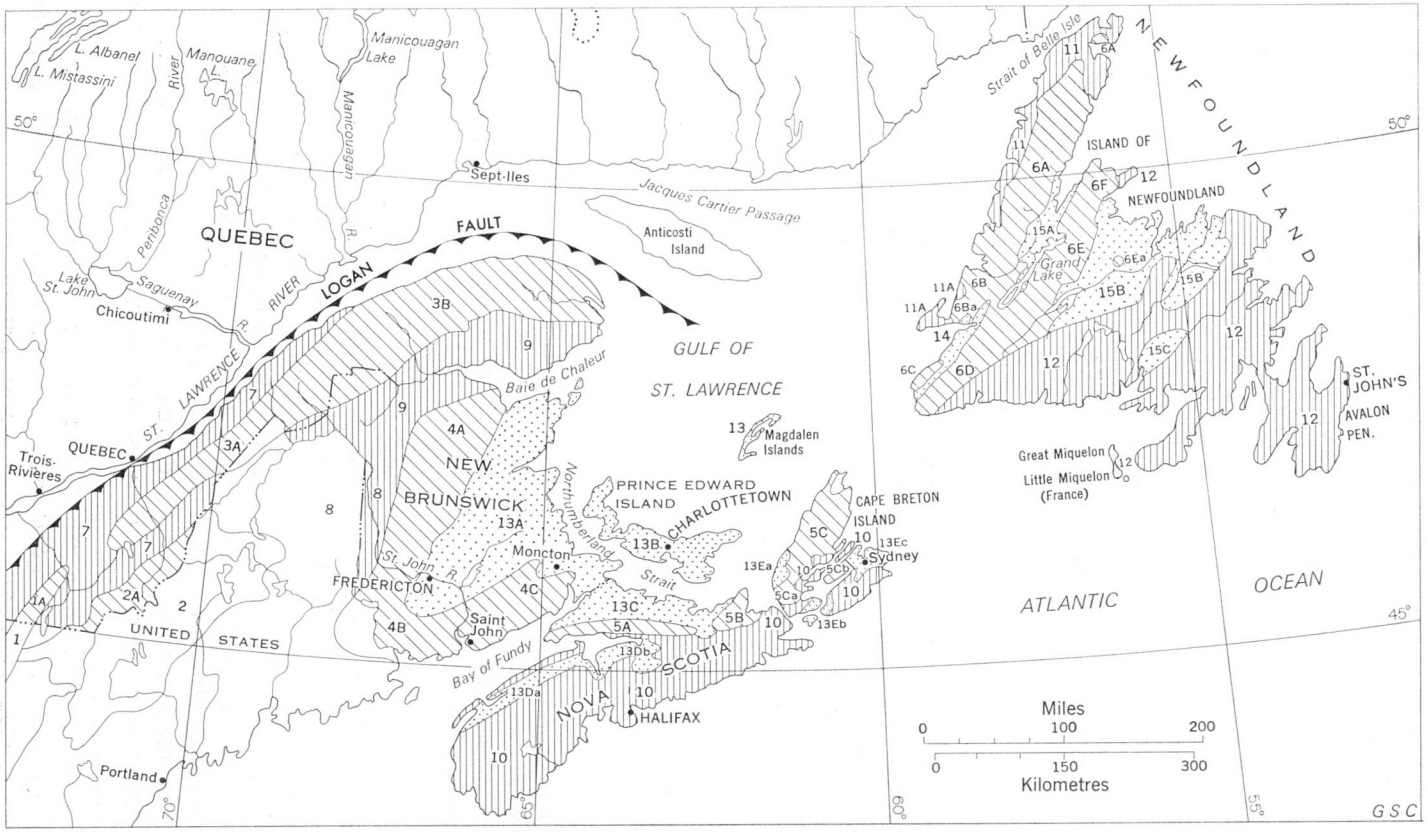


FIGURE 28. Physiographic divisions of the Appalachian hydrogeological region.

McAlary (1960) described the groundwater supplies in the Moncton basin of New Brunswick. In 1964 the author described the geology and hydrogeology of the Moncton map-area.

Brandon completed three water probability maps of the Canso, Truro east half and Halifax east half, and Truro west half map-areas of central Nova Scotia. He also wrote a paper (1966) on the groundwater hydrology and water supply of Prince Edward Island.

Parsons (*in press*) compiled two water probability maps of the Amherst and Fredericton areas of New Brunswick. Mossman (1964) wrote two internal reports for the Nova Scotia Department of Mines concerning the Pleistocene geology and groundwater chemistry of the northern part of the Annapolis Valley.

Hydrogeology

Groundwater movement in the consolidated rocks occurs through discontinuities of the rocks—that is, joints, fractures, bedding planes, and faults. Relatively little groundwater moves through the pore spaces. Thus, to depict the water-transmitting properties, the rocks should be grouped according to the dimensions of their discontinuities. These discontinuities are often hard to locate, for they are concealed by till. Thus it is very difficult to make such a map. However, a map (Fig. 27) was made of the bedrock, and it showed the different hydrostratigraphic units in this region. A hydrostratigraphic unit is defined here as a sequence of rocks (formations, groups, or systems) that yield approximately the same quantity and quality of groundwater.

This map shows only the Maritimes—that is Nova Scotia, Prince Edward Island, the Magdalen Islands, and New Brunswick. Lack of enough data precludes construction of similar maps of Newfoundland and the Quebec Appalachians. The descriptions of these units are interrelated, and specific information is given only for areas where detailed work has been done. There will always be local exceptions to the descriptions given here, but these descriptions are believed to be generally true and give the regional aquifer potential of the unit.

Maritimes

In the Maritimes, one of the more distinctive hydrostratigraphic units, the Windsor Group (3), comprises limestones, gypsum, anhydrite, salt, and red beds. The presence of the evaporites in this unit results in groundwater with a high total dissolved solids (tds), which may be either chloride or sulphate or both. Although this group is thin in New Brunswick, its presence is important since the contaminating effect on the groundwater is present farther down the flow system.

The next most distinctive unit, the Basement Complex (1), comprises dominantly igneous and metamorphic rocks of pre-Carboniferous age, which yield relatively little groundwater.

The rocks between these two distinct units were grouped together to form hydrostratigraphic unit 2, the pre-Windsor sediments, which comprise conglomerates, sandstones, siltstones, and shales of mainly Mississippian age.

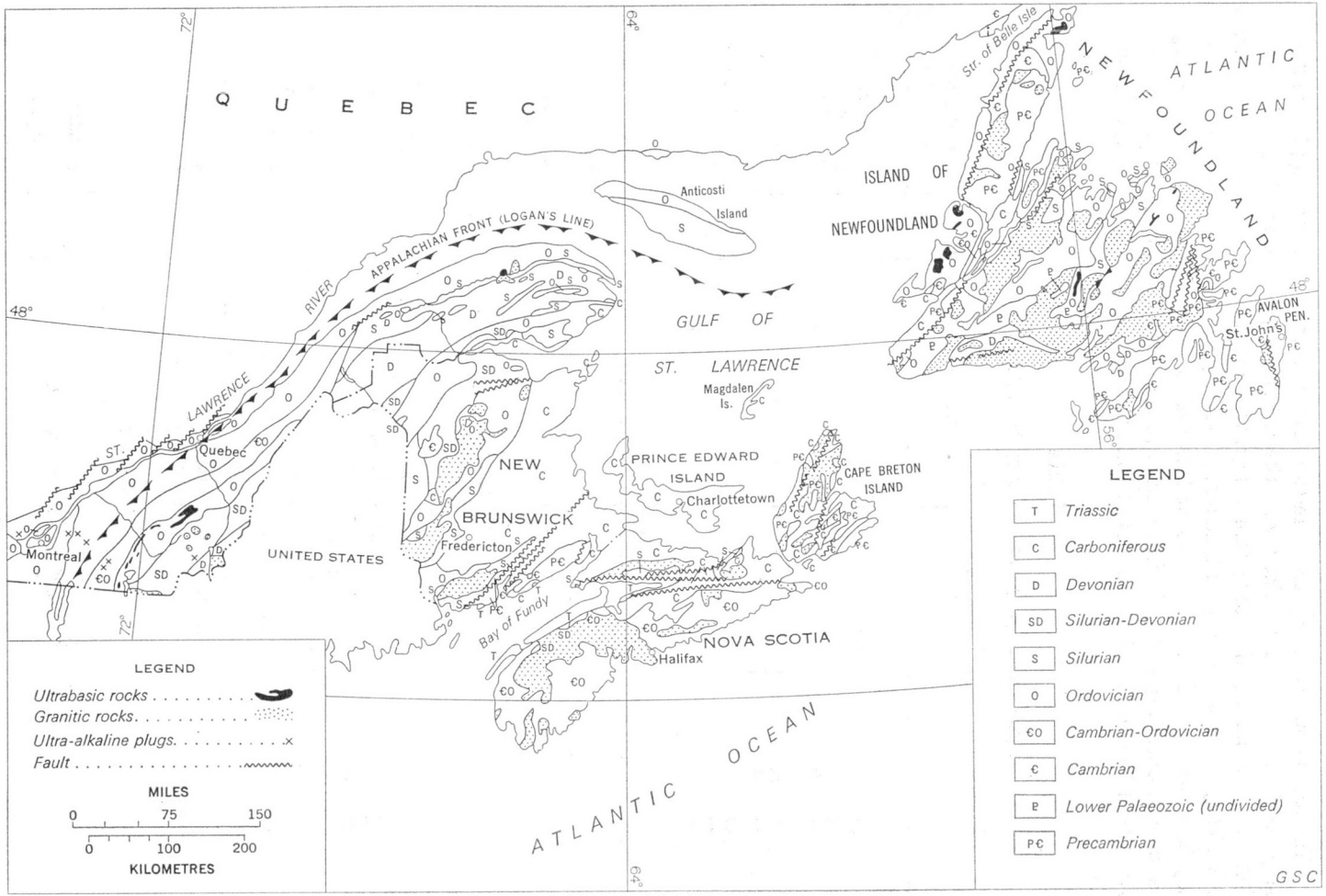


FIGURE 29 Generalized geology of the Appalachian hydrogeological region.

The sediments overlying the Windsor Group form hydrostratigraphic unit 4, the post-Windsor sediments. They comprise sandstones, conglomerates, shales, and siltstones, dominantly of Pennsylvanian age, with some Mississippian and Permian rocks. They are more permeable than similar rock types of unit 2, for, unlike unit 2, they have not been deformed or so well indurated. Where exposed to weathering, the rocks of unit 4 are soft and friable and thus permeable. This unit is situated in the lowlands of New Brunswick and Nova Scotia and extends over all of Prince Edward Island.

Table V

Hydrostratigraphic Units of the Maritime Provinces

	System or epoch	Hydrostratigraphic Unit		Dominant Lithology
UNCONSOLIDATED ROCKS	Recent		7	beach sands tidal muds river alluvium
	Pleistocene		6B 6A	sand and gravel till
	UNCONFORMITY			
CONSOLIDATED ROCKS	Triassic		5B 5A	basalt sandstone
	UNCONFORMITY			
	Permian	post-Windsor sediments		sandstone claystone
	Pennsylvanian			
	Mississippian			
	Mississippian	Windsor Group	3	limestone gypsum salt
	Mississippian	pre-Windsor sediments	2	sandstone shale
	UNCONFORMITY			
	pre-Carboniferous	Basement Complex	1	granites slates argillites

Unit 5 comprises the Triassic rocks that are adjacent to the Bay of Fundy. These rocks comprise two distinct types—basalt that contains very little groundwater, and the sandstone that is one of the better aquifers in Nova Scotia.

The remaining hydrostratigraphic units are confined to the surficial deposits and are not shown in Figure 27. The best aquifers are the sand and gravel deposits of the Pleistocene and Recent epochs, but little detailed information is available concerning their size and location. Thus a map similar to Figure 27 cannot be made at this time; Figure 30 (*in pocket*), however, shows the general areas where sand and gravel deposits are known to be extensive.

Consolidated Rocks

Basement Complex

The Basement Complex (hydrostratigraphic unit 1) comprises all the rocks in the Maritimes ranging from Precambrian to Devonian age. These rocks are igneous and metamorphic, and mainly comprise granites, slates, argillites, quartzites, and a few sediments of limestone and shale. Nearly all of them were deformed by the Acadian orogeny at the end of the Devonian period, and some were deformed by the Taconic orogeny in Middle and Late Ordovician time. These disturbances, however, do not appear to have developed the large interconnected fracture zones associated with good aquifers. For that reason, numerous lakes have developed on much of the Devonian granite of New Brunswick and Nova Scotia and on the rocks of the Meguma Series of Nova Scotia (Geol. Surv. Can. 1949). This unit underlies about half the Maritimes. The rocks are resistant to erosion and hence form the highlands and uplands (Fig. 28) of New Brunswick and Nova Scotia.

In the Moncton area of New Brunswick very small flows (up to 10 barrels a day) of salty water were reported from depths varying between 393 feet to 1,045 feet. For the Amherst area of New Brunswick, Parsons (*in press*) reported yields up to 4 gpm from these rocks and states that significantly higher yields cannot be expected. Numerous small springs provide domestic supplies in most of this area.

In Nova Scotia, estimated yields (from flow in fractures and joints in this unit) of more than 20 gpm are unusual and yields of 1 to 5 gpm are commonly to be expected. Deep drilling—that is, drilling to a depth of more than 200 feet—is not recommended, for the joints and fractures usually close with depth. The better wells occur in valleys at shallower depths.

The groundwater is a calcium bicarbonate type generally containing 50 to 150 ppm as total dissolved solids. Iron and manganese concentrations are locally excessive, especially in the Meguma Series of Nova Scotia.

Pre-Windsor Sediments

The pre-Windsor sediments (hydrostratigraphic unit 2) comprise all those Mississippian rocks that overlie the Basement Complex and underlie the Windsor rocks. They comprise mainly conglomerates, sandstones, siltstones, and shales. Usually these rocks have been folded and are more compact than the post-Windsor sediments. This compaction limits the permeability of the rock and hence the yield.

This unit is confined to the central part of the Maritimes and occurs mainly in upland areas. In the Moncton area, the unit is a poor aquifer, yielding between 3 and 10 gpm. In the Amherst area the common yield is 10 gpm and in more favourable locations (i.e., some valleys) is up to 50 gpm. In Nova Scotia the yield varies between 5 and 25 gpm.

Except where the groundwater has flowed through Windsor sediments before entering this unit, the quality of the groundwater is good. It is predominantly a calcium bicarbonate type containing 100 to 300 ppm of tds. Water of this composition is medium-hard to hard. Some salt springs occur in the Albert Formation of New Brunswick, but the salt is generally well buried and the salt springs are decidedly local.

Windsor Group

The Windsor Group (hydrostratigraphic unit 3) comprises marine limestone, dolomite, anhydrite, gypsum, salt, shale, and siltstone of Mississippian age. These have been folded and faulted. Along the faults and joints, groundwater has dissolved the rock and developed solution channelling in the more soluble rocks.

Although most extensive in central Nova Scotia, this unit extends into New Brunswick, where it occurs as thin beds traceable over long distances. It is also present on the Magdalen Islands.

The presence of solution channels causes the yield to be extremely variable.

The water is excessively hard or saline, or both, often containing more than 10,000 ppm as tds. In some areas wells have been drilled through these rocks, the zone with poor quality water then cased off, and good-quality water has been obtained from underlying rocks. The main constituents in the groundwater are sodium, calcium, chloride, and sulphate.

Post-Windsor Sediments

The post-Windsor sediments (hydrostratigraphic unit 4) comprise the Carboniferous and Permian sediments overlying the Windsor Group. They are composed mainly of sandstone and claystone, with some conglomerate and siltstone. These rocks are dominantly of Pennsylvanian age, but some are Mississippian and some Permian. Included in this unit are the Hopewell and Petitcodiac Groups of the Moncton basin, the rocks of the New Brunswick Shelf, and the rocks of the Canso, Riversdale, Cumberland, and Pictou Series of Nova Scotia. Except for some of the Mississippian and the lower parts of the Pennsylvanian system, the rocks are undisturbed and nearly flat lying. These rocks are not well indurated, and where they have been exposed to weathering they decompose rapidly. Thus, because of their greater permeability, it is to be expected that the rate of infiltration of precipitation into these rocks is higher than in the previous units.

The unit covers about one third of the Maritimes and is situated in the lowlands of the Gulf of St. Lawrence Plain (*see* Fig. 27 *in pocket*, and Fig. 28).

In the Moncton area of New Brunswick the Hopewell Group is an aquitard. The basal conglomerate is poorly sorted and contains a considerable amount of shale

and claystone matrix between the granite pebbles. In the Amherst area the yield varies up to 10 gpm.

The Petitcodiac Group varies from a poor to a good aquifer. In the Moncton basin it has been subdivided into the Boss Point, Salisbury, and Richibucto Formations (Carr, 1964). The Boss Point Formation is the best aquifer within the Moncton basin. McAlary (1960) reported a flow of 1.2×10^6 gpd of fresh water coming from 40 feet below the top of the formation in Imperial Oil's Pollett River number 1 and a flow of 1×10^5 gpd of fresh water from 19 feet below the top of the formation in New Brunswick Oilfields Turtle Creek number 1. A spring 10 miles northeast of Sussex flows continuously at about 1×10^6 gpd. Five wells 3 miles northwest of Sackville, New Brunswick, produce 300 to 600 gpm from 90 to 150 feet within the Boss Point Formation.

The Salisbury Formation yields about 10 gpm in the Moncton area, whereas it yields from 5 to 50 gpm in the Petitcodiac area. Two values of permeability[1] of this formation in the Moncton area were determined to be 2.2 and 5.3 gallons per day per square foot (gpd/ft^2) and the corresponding values of storage coefficients were 2.2×10^{-3} and 4.0×10^{-4} (Carr, 1964).

The Richibucto Formation usually yields at least 20 gpm. One of the town wells at Shediac has yielded more than 600 gpm. Values of permeability varied from 2.1 to 195.0 gpm/ft^2, and values of storage coefficients varied from 2.2×10^{-3} to 4.0×10^{-4}.

Elsewhere in the areas underlain by the Pennsylvanian sediments of New Brunswick, yields of 25 gpm can generally be obtained. At Fredericton four wells drilled in these rocks to depths between 187 and 400 feet yield between 150 and 500 gpm (Parsons, *in press*).

In Nova Scotia, rocks of the Canso and Riversdale Series may yield up to 200 gpm. The rocks of the Cumberland and Pictou Series, north of the Cobequid Mountains, have higher yields than the rocks of these series south of the complex, where they are more consolidated. At Amherst a well 700 feet deep produces 670 gpm. Well yields north of the complex generally vary between 10 and 75 gpm, whereas well yields south of the Cobequid Mountains vary between 5 and 25 gpm.

Groundwater from this hydrostratigraphic unit is of good quality and generally contains from 100 to 300 ppm of tds. Springs and shallow wells may normally have less than 100 ppm of tds: the water is a calcium bicarbonate type. A few isolated occurrences of saline water do exist in these rocks, the groundwater having presumably come into contact with salt from the Windsor Group.

Some wells adjacent to the coast in Prince Edward Island and at Shippegan, New Brunswick, have been overpumped and have caused sea water to intrude landward.

Triassic Rocks

The Triassic rocks (hydrostratigraphic unit 5) comprise sandstone and shale of the Annapolis Formation (subunit 5A) and the basalt and diabase (subunit 5B). These rocks are essentially flat lying and undisturbed. They are situated adjacent to

[1]All values of permeability are field coefficients of permeability as defined by Wenzel, 1942, p. 7.

the Bay of Fundy in Nova Scotia, the basalt forming the upland of North Mountain and the sandstone and shale forming the lowland of the Annapolis Valley and Minas Basin. Some isolated occurrences of Triassic rocks are present on the south shore of New Brunswick and in Chedabucto Bay, Nova Scotia.

Groundwater movement through the Annapolis Formation occurs along joints and bedding planes, but intergranular permeability is present and appreciable. Aquifer tests in the vicinity of Debert indicated a range of permeability from 35 to 170 gpd/ft^2. The highest reported yield from the formation is 400 gpm, from one well in Truro. This is probably the best bedrock aquifer in Nova Scotia. The basalt and diabase generally yields less than 3 gpm (Parsons, *in press*).

Groundwater from this unit contains between 50 and 200 ppm of tds. At Kingsport, Nova Scotia, sea-water intrusion has occurred and a sample of groundwater contained 1,400 ppm of tds, mainly sodium chloride.

Surficial Materials

Pleistocene Deposits

The Pleistocene deposits (hydrostratigraphic unit 6) comprise till (subunit 6A) and sand and gravel deposits (subunit 6B). Most of the Maritimes is covered with a thin layer of till, which is a poor aquifer, and the surface run-off from it is probably high. Most shallow dug wells produce fairly soft water having less than 70 ppm of tds.

The Pleistocene sand and gravel deposits (subunit 6B) occur as eskers, kames, and outwash plains. Figure 30 (*in pocket*) illustrates the areas where these deposits are most extensive. They are not widespread or abundant, but the rainfall is so high in this region and the deposits so permeable that they are the most important aquifers in the Appalachian region.

The location and size of these aquifers is not well known. The larger deposits, however, occur in the Annapolis Valley of Nova Scotia, in the Saint John River valley between Keswick and Belleisle Bay and the Sussex area of New Brunswick, and on the west coast of Newfoundland. These deposits are variable in thickness and permeability; hence their yield is unknown.

Drilling by the Nova Scotia Department of Mines has indicated up to 160 feet of glacial sediments in the Annapolis Valley (Mossman, 1964). These glacial sediments comprise two gravel beds, one 95 feet thick and the other 20 feet thick, occurring between clay and silt. They are situated a mile south of Waterville in a postulated buried channel, which extends from Berwick to 2 miles north of Kentville. Pumping tests indicate that wells in this aquifer could possibly yield in excess of 1,000 gpm (J. F. Jones pers. com.)

In New Brunswick, the town of Sussex obtains its water supply from a well 115 feet deep, drilled in sand and gravel. The well is rated at 500 gpm with a specific capacity of 17 gpm per foot of drawdown (12-inch-diameter well). Three municipal wells at Fredericton produce water from sand and gravel adjacent to the Saint John River from depths of 113 feet to 146 feet. Their initial capacities were between 590 and 1,125 gpm and their specific capacities yield between 32 and 165 gpm per foot of

drawdown (12- to 18-inch-diameter wells). At St. George, a well drilled 56 feet in sand and gravel adjacent to the Magaguadavic River initially yielded 700 gpm. The specific capacity of this well is 233 gpm per foot of drawdown (diameter of well not recorded). Three miles north of St. Stephen, a sand and gravel aquifer adjacent to Dennis Stream yields 2 million gpd.

Groundwater from these outwash deposits generally contains less than 200 ppm of tds and is normally a calcium bicarbonate type.

These deposits are often confined to valleys. Thus they are in a position to intercept a maximum amount of groundwater discharge. Also, they are adjacent to rivers from which surface water can be induced by pumping. Thus these aquifers have the potential for yielding more water than any other aquifer in the Maritimes.

Recent Deposits

The deposits of the Recent epoch (hydrostratigraphic unit 7) comprise mainly beach sands, tidal muds, and river alluvium.

The beach sands are especially abundant on the north shore of Prince Edward Island and the east shore of New Brunswick. In both areas beaches and sand spits are well developed, but very little fresh water can be pumped from these sands until salt water is induced inland.

Alluvium of sand and gravel occurs in valleys of some of the larger rivers, such as the Saint John (Fig. 30). It is usually derived from Pleistocene glaciofluvial material and occurs as meander bars and terraces. These deposits can yield a great deal of water by induced infiltration.

The tidal muds are situated in the estuaries of rivers entering the Bay of Fundy from Nova Scotia and New Brunswick. These muds, having been deposited under salt-water conditions, have a low permeability, and contain saline water. Fresh water, however, can sometimes be obtained from the underlying bedrock. Flowing wells in the lowlands between Fort Lawrence and Amherst produce potable water from a depth of 100 to 150 feet. The water here is a sodium calcium bicarbonate type with 150 ppm of tds. Several occurrences of saline water have been reported in the Tantramar Flats, east of Sackville, down to a depth of 500 feet (Parsons, *in press*).

Quebec and Gaspé

No published information concerning the groundwater of this area is available, but certain inferences can be made from the known topography and geology.

This area, especially the Gaspé Peninsula, is distinctly rugged. The highest mountains, the Shickshocks, exceed 4,100 feet in places. Parts of this range are high enough to be underlain by permafrost (McGerrigle, 1954, p. 16). The mountain tops present a flat surface and are part of the general peneplain, which slopes from the Notre Dame Mountains southeasterly to the Atlantic. Deep dissection of the plateau surface by numerous rivers has, however, produced an extremely rugged topography. The river valleys are mainly steep-sided and narrow, and their bottoms are in many

places 1,000 feet or more below the neighbouring uplands (Dresser and Dennis, 1944). Such deep valleys will expose numerous springlines. Thus a great deal of the groundwater will not travel far before becoming surface water.

Consolidated Rocks

Most of the rocks are Palaeozoic and have a dominant northeasterly trend, which swings eastward in the Gaspé Peninsula. The rocks of the Quebec Appalachians can be divided into three major structural belts—northern, central and southern (Geol. Surv. Can., 1958).

The northern belt is 10 to 25 miles wide and extends the full length of the Quebec Appalachians. Its northern boundary is marked by Logan's Line. The rocks are mainly slates of the Lower and Middle Ordovician system that were tightly folded and fractured in the Taconic orogeny (Neale, *and others*, 1961).

The central belt comprises two structurally distinct regions, one in the Gaspé Peninsula and the other in the Eastern Townships. In the Gaspé Peninsula the belt is underlain mainly by Silurian and Devonian rocks, dominantly the Gaspé sandstone and the Gaspé limestone. This area was folded in the Acadian orogeny into a broad synclinorium having a general plunge to the east.

In the Eastern Townships this central belt is underlain mainly by Cambrian and Ordovician rocks composed of feldspathic sandstones, greywacke, volcanics, and schists. These rocks were folded in the Taconic orogeny and refolded in the Acadian orogeny. Thus the rocks are very tightly compressed. Physiographically this belt is expressed by the Sutton and Notre Dame Mountains and part of the Eastern Quebec Uplands (Fig. 28). A small area of rocks on the north shore of Chaleur Bay has had a similar structural history.

The southern structural belt is 25 to 30 miles wide and comprises rocks mainly of Ordovician to Devonian age that were folded by the Acadian orogeny. These rocks are sediments with some volcanics, mainly shale, limestone, and quartzites. Small basins have developed on these rocks at the eastern end of the Gaspé Peninsula, and in them Carboniferous sandstones and shales were deposited. These rocks, however, are limited in their areal extent and are seldom found more than a few miles inland from the coast.

It is likely that large quantities of groundwater can be obtained from the limestone and sandstone of the central belt. Cumming (1959, p. 35) reported that the beds in the lower part of the Gaspé limestone have appreciable porosity and permeability. Also, where the fossiliferous part of the limestone has been exposed to weathering, good porosity has developed. Parks (1930, p. 40) considered that the Gaspé sandstones and conglomerates had sufficient porosity to retain large quantities of petroleum; thus these rocks are probably good aquifers too. Also large quantities of groundwater can probably be obtained from the Carboniferous sandstones despite their limited areal distribution.

Analysis of 21 pumping tests supplied by the Quebec Department of Natural Resources indicates that wells drilled in the bedrock to a depth of 135 feet yield about 20 gpm. Occasionally yields as high as 45 and 65 gpm have been encountered.

Surficial Materials

The aquifers yielding the greatest amount of groundwater are the Pleistocene sand and gravel deposits. Probably the most utilized sand aquifers are the marine terraces. These occur as a discontinuous group around the Gaspé coast (Fig. 30). They form most of the good farmland in northern Gaspé. Often the terraces can be traced up the river valleys. The marine terraces are confined to the neighbourhood of the sea and never rise more than 200 or 300 feet above the present sea-level.

Inland the glacial drift is thin and discontinuous, but sand and gravel deposits, although not abundant, are present as kames, eskers, and river terraces. These deposits were formed as the sand and gravel were washed out under and in front of the thinning ice masses. They are confined mainly to river valleys such as the Restigouche and the Matapedia. The Matapedia Valley contains numerous deposits of stratified gravels, especially between Amqui and the south end of Matapedia Lake (Alcock, 1935). In the lower parts of some of the larger streams, river terraces have developed in the unconsolidated material of the valley bottoms. On the Matapedia, river terraces occur at elevations up to 500 feet.

In the Beauceville area, Gadd (1964) reports three main types of sand and gravel deposits—buried channel deposits, glacial outwash, and alluvium. The buried channel deposits comprise well-sorted bedded sand and gravel occurring between two tills. This type of deposit is confined mainly to the Chaudière and Linière Valleys, where it attains a maximum thickness of 115 feet. The glacial outwash occurs as kame moraines, kame terraces, outwash plains, and eskers, all related to the last ice recession. They are developed in the valleys and on the slopes. The alluvium comprises coarse gravel and sand occurring as bars and spits in the Chaudière Valley and its tributaries.

Lee (1962) has reported two types of sand and gravel deposits in the Rivière du Loup area that should be good aquifers. First are the outwash deposits comprising coarse silty gravels and silty sands; these occur at elevations above 350 feet. Second are the high- and low-terrace sands. The high-terrace sands are clean and gravelly and occur between 50 and 350 feet above sea-level, whereas the low-terrace sands comprise silty sand with mica flakes and occur below an elevation of 50 feet. These terrace sands are from 2 to 20 feet thick. In the Trois Pistoles area, Lee reported esker sands occurring in the uplands along with extensive outwash deposits of gravelly sands, sands, and silty sands. These are proglacial facies of deltaic sands. They range in thickness from 15 to 45 feet. The high-terrace sands also occur in this region.

Analysis of eight pump tests indicates that wells drilled to 55 feet in the more permeable surficial material yield 60 gpm. Occasionally yields as high as 110 and 136 gpm have been encountered.

Newfoundland

No groundwater work has been done here to date. Inspection of the topographic map of the Island, however, reveals numerous ponds, swamps, peat bogs, and lakes, the result of a poor drainage system. Poorly drained lands have been mapped in all parts of Newfoundland, and a much higher percentage of the Island is poorly drained

than well drained (Mollard and Munn, 1955, p. 362). A well-developed drainage system probably existed before Pleistocene time; the glacial drift, however, has disrupted the system. Evidently the climate is conducive to the growth of moss and peat, which has developed throughout the Island. These peat and moss areas retain the surface water and permit it to drain off slowly. Nearly all the municipalities obtain their water supply from surface water sources.

Consolidated Rocks

Examination of the geological map (Geol. Surv. Can., 1955) reveals that most of the Island is underlain by igneous and metamorphic rocks and some well-indurated sediments, all of a pre-Carboniferous age. Where the bedrock is well exposed at the surface, numerous joints and faults can be seen in certain aerial photographs. Often a line of trees or shrubs will occur along these joints, thus attesting to their ability to collect soil and water. It is reasonable, however, to assume that the abundance and extent of these discontinuities will be extremely variable. In the Great Northern Peninsula, near Hawke Bay, sinkholes are well developed in the areas underlain by Cambrian and Ordovician limestone.

The Carboniferous sediments are situated in the western part of the Island, mainly in the Anguille Mountains, the St. George's Lowland, and the Grand Lake Lowlands. The Mississippian rocks comprise sandstone and shale mainly, but the Codroy Group contains siltstone, limestone, and gypsum. Locally a karst terrain has developed on the gypsum (Hayes and Johnson, 1938; McKillop, 1963); no salt, however, has been found. The Pennsylvanian rocks consist mainly of sandstone, shale, conglomerate, and a little coal. These rocks are non-marine and seem to be similar in many respects to the Petitcodiac Group of New Brunswick and probably have a similar water-transmitting ability.

Surficial Materials

The surficial deposits of the Pleistocene epoch can be grouped into four distinct geographical regions—the Avalon Peninsula, the outer drift zone, the inner drift zone, and the western half of the Island (Fig. 30).

On the Avalon Peninsula ice movement was coastward from the centres of each of its elongated peninsulas (Jenness, 1960). The local groundwater supply comes from the ground moraine, which is generally shallow. Larger supplies are obtained from surface water. There are only a few isolated sand and gravel outwash deposits, which may be good aquifers. These are usually situated near the coast (Henderson, 1960).

The glacial deposits of eastern Newfoundland west of the Avalon Peninsula fall into two groups—the outer drift zone and the inner drift zone. The outer drift zone comprises the area between the discontinuous end moraine (Fig. 30) and the Avalon Peninsula. The deposits there are ground moraine, drift ridges, eskers, kame terraces, and deltas. Local supplies can probably be obtained from the ground moraine, but the most important aquifers are the outwash deposits. They radiate from the

margin of the inner drift zone outward towards the coast-line down river valleys and terminate at the coast as deltas. In Bonavista Bay, at Eastport and Sandy Cove (a mile to the southeast), the deltaic sediments have formed two 100-foot cliffs. These deposits should be excellent aquifers. The glacial deposits of the inner drift zone are ground moraine, eskers, and a few kames and kame terraces. The eskers form a broad belt 20 to 30 miles wide just inside the end moraine. Locally these eskers may be good aquifers if they are situated in a favourable topographic position.

Glacial deposits of the western half of Newfoundland are grouped together, since little is known about them except on the coast-line. Extensive outwash deposits and uplifted marine terraces occur on the west side of the Island.

At the mouth of almost every stream on the west coast there is a flat-topped alluvial deposit consisting of coarse and fine sediments. At many places on each side of the delta flat a terrace continues, winding in and out of the re-entrants of the coast. These deposits are generally excellent aquifers. The wells of very large capacity at Stephenville are giving groundwater from these deposits.

Morainal and glaciofluvial deposits in the St. George's Lowland, reach a thickness of more than 100 feet in many places (Hayes and Johnson, 1938). North of the Anguille Mountains, outwash sands at the mouths of the larger streams form higher headlands than the bedrock. At Robinson's Head the wave-cut bench is about 50 feet above sea-level, and the bedded unconsolidated sands are more than 200 feet thick above it. These should be excellent aquifers.

In the Great Northern Peninsula the marine sediments, comprising sand and well-rounded gravels, extend over a large area (Fig. 30). Much of this, however, is overlain by peat bogs (Mollard and Munn, 1955).

Large deposits of stratified sand and gravel occurring as terraces are situated along Indian River.

Of the total area of Newfoundland, 19,934 square miles, only 715 square miles, about 3.6 per cent, is underlain by large potential sand and gravel aquifers.

Climate

A humid continental climate prevails over all parts of the Atlantic Provinces except the south shore and the Great Peninsula of Newfoundland, where a subarctic climate exists.

The region is situated near where practically all the cyclonic storms of North America leave the continent. These storms control the weather, for in this region the warm moisture-laden air masses from southern latitudes are constantly meeting cold air masses from the northern interior. In summer these storms pass to the north, drawing in warm air from the south and southwest; in winter they pass along the southern border inducing invasions of cold air from the interior (Putnam, 1940).

Inland parts of this region are shut off from the tempering influence of the ocean. Thus, at places like Grand Falls, New Brunswick, the monthly temperatures are lower in winter and higher in summer than at Saint John or Halifax. At these latter places precipitation and humidity are higher.

Temperature

The mean annual minimum temperature ranges from 0°F at Yarmouth to minus 30°F in central New Brunswick, and the mean annual maximum temperature ranges from 75°F in the Great Northern Peninsula of Newfoundland to 90°F in central New Brunswick.

The number of frost-free days in the inland areas is about 100, whereas along the coast, at Halifax and Saint John, it is about 170. In highly elevated inland areas the number of frost-free days may be around 60.

Precipitation

In the Atlantic Provinces the mean annual precipitation varies from 57 inches a year on the east coast of Nova Scotia and the north coast of Newfoundland to 34 inches a year in the northern parts of New Brunswick and Newfoundland. In the Gaspé Peninsula the precipitation varies from 36 to 40 inches a year. The precipitation is fairly well distributed throughout the year, and about a third of it falls during the growing season.

The winters in northern New Brunswick are long and cold, with more than 100 inches of snow. The winters in the Bay of Fundy and the Annapolis Valley region are milder, with 70 to 80 inches of snow. The relative humidity is high along the Atlantic coast, where it ranges between 85 and 90 per cent; in the interior of New Brunswick it falls below 75 per cent.

Potential Evaporation

Potential evaporation is the amount of water that would be transferred from the soil to the atmosphere by evaporation and transpiration if it were constantly available in optimum quantity.

In Prince Edward Island evaporation has been measured at the Charlottetown Experimental Farm with a black Bellani plate. For the summer months of 1960-63 a range of 9 to 13 inches was obtained. The potential evaporation computed by the Thornthwaite method for the same four months yielded a value of 16.7 inches. Brandon (1966) concluded that the actual annual evaporation ranges from about 12 to 19 inches.

In Nova Scotia the potential evaporation (Table VI), as determined by the Thornthwaite method, ranges from 18 to 26 inches annually.

Table VI

Potential Evaporation in Nova Scotia[1]

Location	Annual P.E. (inches)
Sydney	20.8
Springfield	18
Yarmouth	21.6
Annapolis Royal	26
Halifax	24.5

[1]N.S. Soil survey Repts. 8, 9, 11, and 12.

Figure 31 shows the precipitation and potential evaporation at Halifax.

At Halifax, precipitation varies from 4 to 5.5 inches a month throughout the year, and only in July do the needs of vegetation rise any higher. Since reserve moisture is stored in the soil, the vegetation does not suffer from any immediate lack. Before this reserve is exhausted, the precipitation rises above the need. For Halifax the total water surplus, the precipitation in excess of need after soil moisture has been replenished, is 32.2 inches (Fig. 31). According to Sanderson (1948) there is no serious depletion of soil moisture during the summer in any part of the Maritimes.

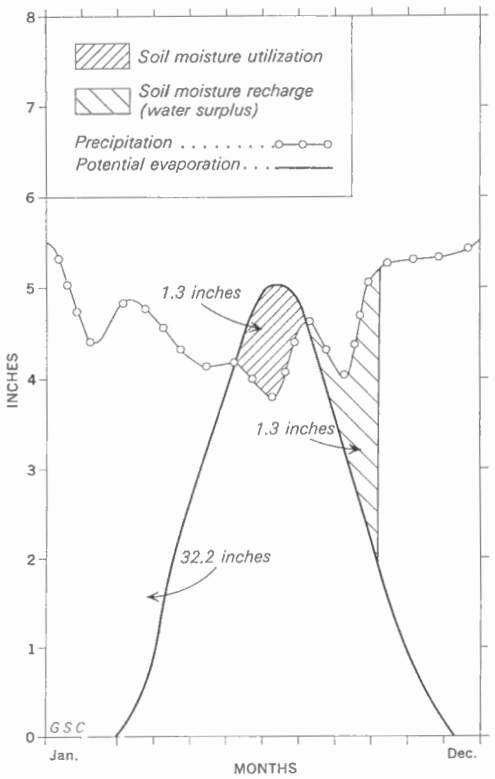


FIGURE 31.
Precipitation and potential evaporation at Halifax, Nova Scotia. Based on 20 years of climatic records (Sanderson, 1948).

Base Flow

Evidence

In Prince Edward Island, Brandon (1966) found field evidence that groundwater is effluent and maintains the flow of rivers. First, there are many springs in depressions occurring throughout the Island. A large number of these springs have flows ranging from 50 to 500 gpm and have formed pools where water bubbles upward, thus indicating upward leakage of groundwater. Secondly, the existence of poor drainage and rich vegetation in valleys is a further indication of groundwater effluence. Farming on the

Island is generally carried out on the high ground and slopes, and not in the valley bottoms, which are swampy in many localities. Thirdly, the cold (44–45°F) groundwater is noticeable in spring-fed streams. Fourthly, springs of fresh water are also noticeable at low tide along river estuaries.

At low flow during the summer of 1964 Parsons (pers. com.) sampled the Kennebecasis River in New Brunswick for detailed chemical analyses to determine whether groundwater was discharging into it.

Sand and gravel occur in the river bed, and underlying this are the nearly flat lying Pennsylvanian sediments and the deformed Mississippian sediments including the evaporites of the Windsor Group (Fig. 32). A hydrograph (Fig. 33) of the river at Apohaqui was plotted. The low flows of July 14, August 14, and September 1 best indicate base flow. Stations A to E (Fig. 32) were sampled on August 14, and the results of the chemical analyses were plotted in Figure 34.

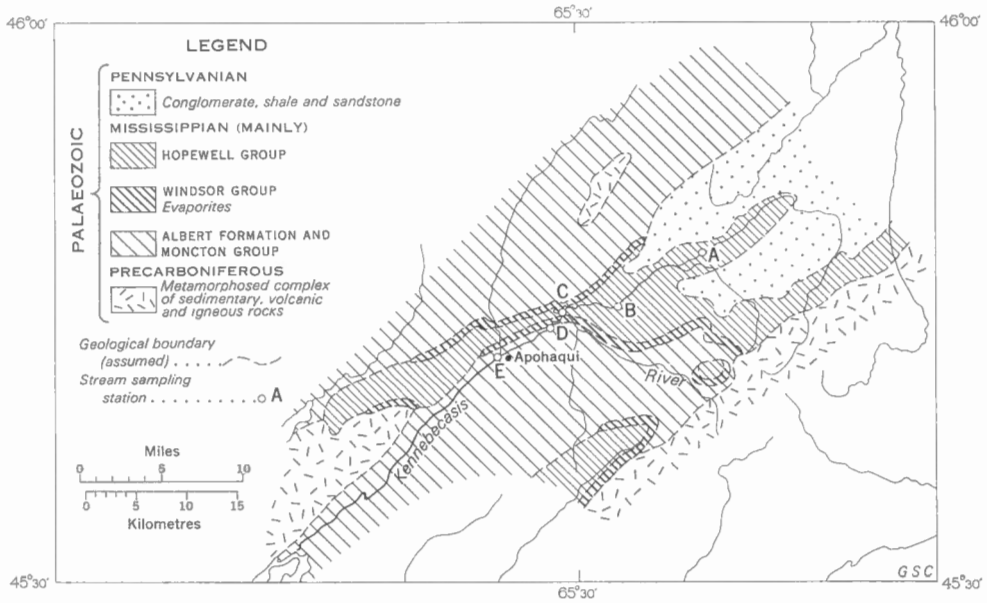


FIGURE 32. Bedrock geology of Kennebecasis Valley, New Brunswick.

At station A the water is a calcium bicarbonate type having a smaller amount of sulphate and an insignificant amount of sodium and chloride. Between stations A and B there is an abrupt increase in the sodium and chloride concentrations, with significant increases in calcium, bicarbonate, and sulphate concentrations. At station C the sodium and chloride concentrations are significantly larger than those at station B. Beyond station C there is a general decrease in the mineralization of the base flow.

It is apparent that groundwater, having come in contact with the Windsor evaporites, contributes significantly to base flow between stations A and B. The high sodium and chloride concentrations at station B indicate the presence of a salt body within the groundwater flow system. Such salt bodies were postulated by Gussow

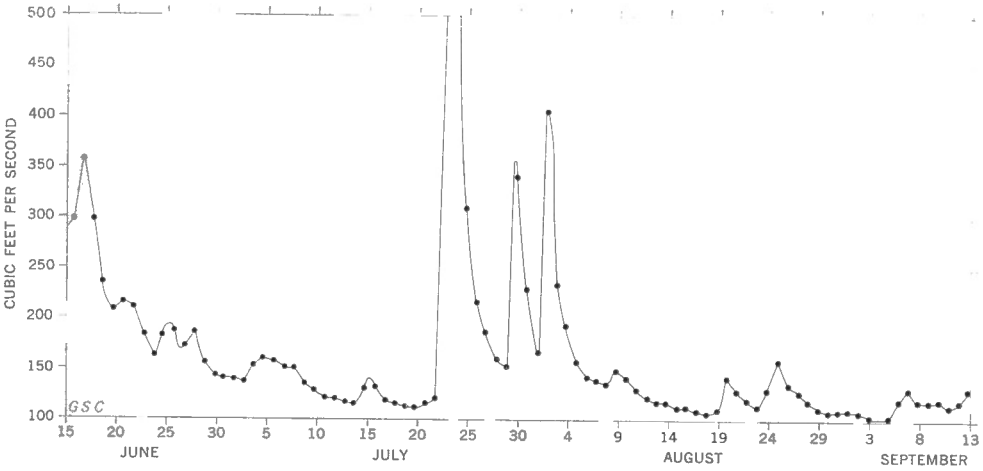


FIGURE 33. Hydrograph of Kennebecasis River at Apohaqui, New Brunswick, 1964.

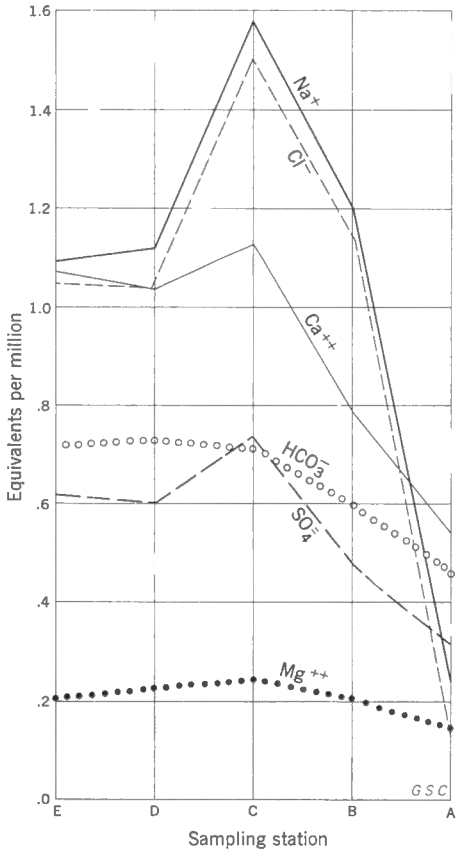


FIGURE 34.
Chemical analyses of Kennebecasis River, New Brunswick, Aug. 14, 1965 (compiled by M. L. Parsons, 1965).

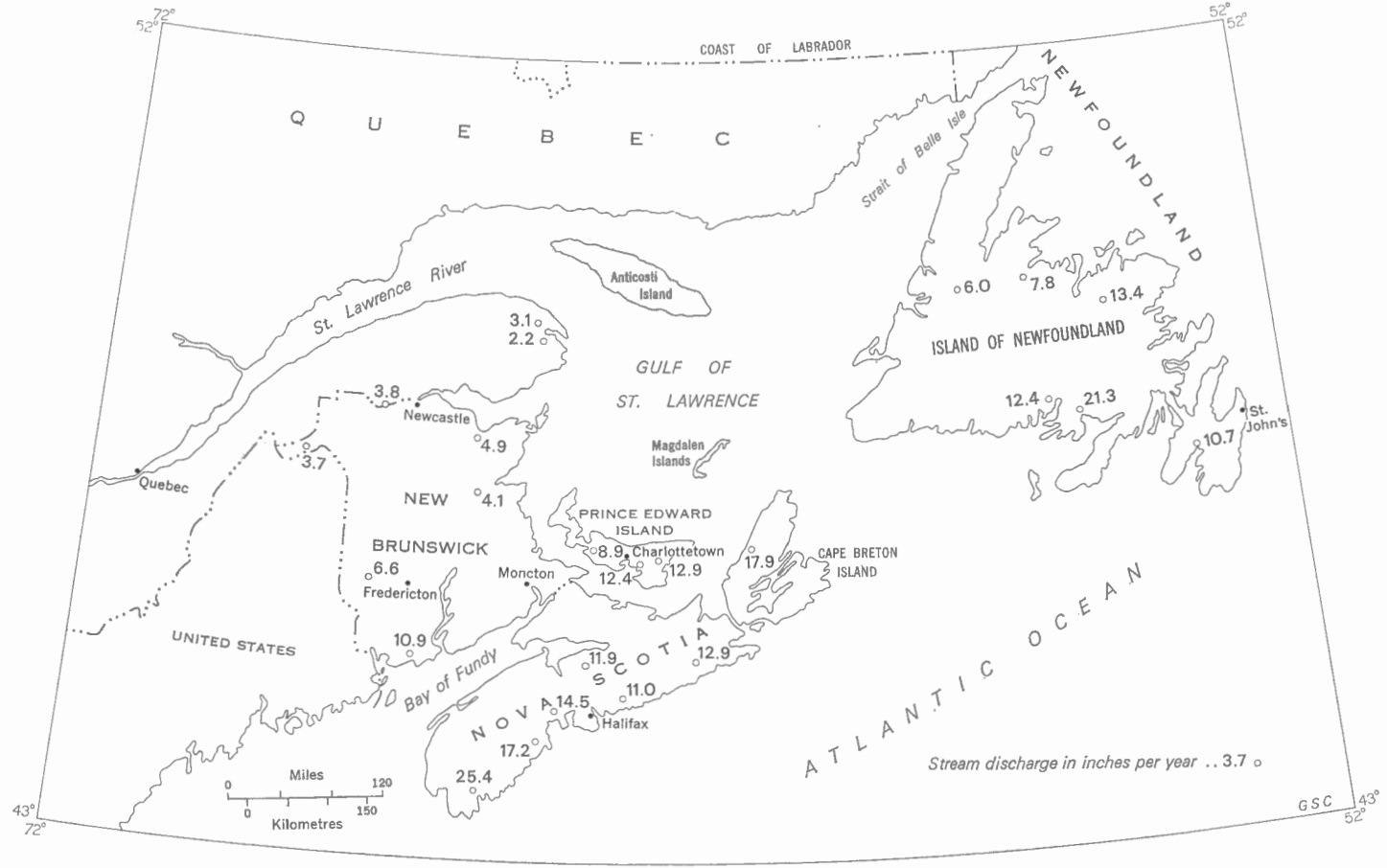


FIGURE 36. Mean minimum winter stream discharge in the Appalachian hydrogeological region.

(1953, p. 1783) to occur along the Anagance anticlinal axis, which coincides here with the Kennebecasis River valley. Also in the vicinity of station B there is a well about 50 feet deep from which salty water flows at the rate of several gallons per minute.

The large sodium and chloride concentrations at station C are evidence that saline groundwater discharge must be contributing to base flow downstream from station B. Between stations C and D there is a decrease in mineralization of the base flow. The decrease coincides with the zero edge of the Windsor Group and continues in a general trend downstream and is probably due to the dilution of base flow by less mineralized groundwater.

Calculations

Lawson has estimated base flow to be 6 inches a year, using a hydrograph considered by Cass-Beggs to be typical of the Maritimes (Fig. 35, *in pocket*).

A map was made to depict base flow (Fig. 36) in the Appalachians. It was based on streamflow data taken from Water Supply Papers of the Department of Northern Affairs and National Resources. Rivers were selected for which nine years of records (1951 to 1960) had been accumulated (for Newfoundland, six years of records—1954 to 1960) and which were believed to be free of regulation. The lowest streamflow, occurring in January, February, or March, is assumed to approximate baseflow, since there are only minor contributions to run-off and no evapotranspiration at this time of year. The data were averaged and converted to inches per year. They represent an approximation of the amount of groundwater contributed to the river between its source and the recording station.

The results obtained ranged from 2.8 to 25.4 inches a year, which is equivalent to a range of 0.2 to 1.9 cubic feet a second per square mile of drainage basin. They are likely too high, for there is some snow melt during these months, especially near the Atlantic Ocean (note how high they are for southern Nova Scotia and Newfoundland) and there may also be some delayed run-off from the numerous lakes and swamps in the region. Thus the results indicate only the order of magnitude and should be regarded as an upper limit of base flow.

Meyboom (1961) calculated the mean minimum stream discharge for Nova Scotia (Fig. 37). He did this by averaging the lowest monthly discharge for each month of the year for a 10-year period (1947–57). The results are in the same order of magnitude as those of Figure 36, although the same period and the same method were not used.

Chemistry

Surface Water

The rivers of this area have been studied by Thomas (1960). Rivers and streams rising in the Gaspé Peninsula—the Matapedia, the Restigouche, and the Madawaska—are harder and more mineralized than the short coastal rivers and those rising farther south in the central plateau of New Brunswick, such as the Miramichi and the Salmon. The rivers of mainland Nova Scotia are even softer than those of New Brunswick, and most have a hardness of less than 10 ppm as calcium carbonate. In Newfoundland the

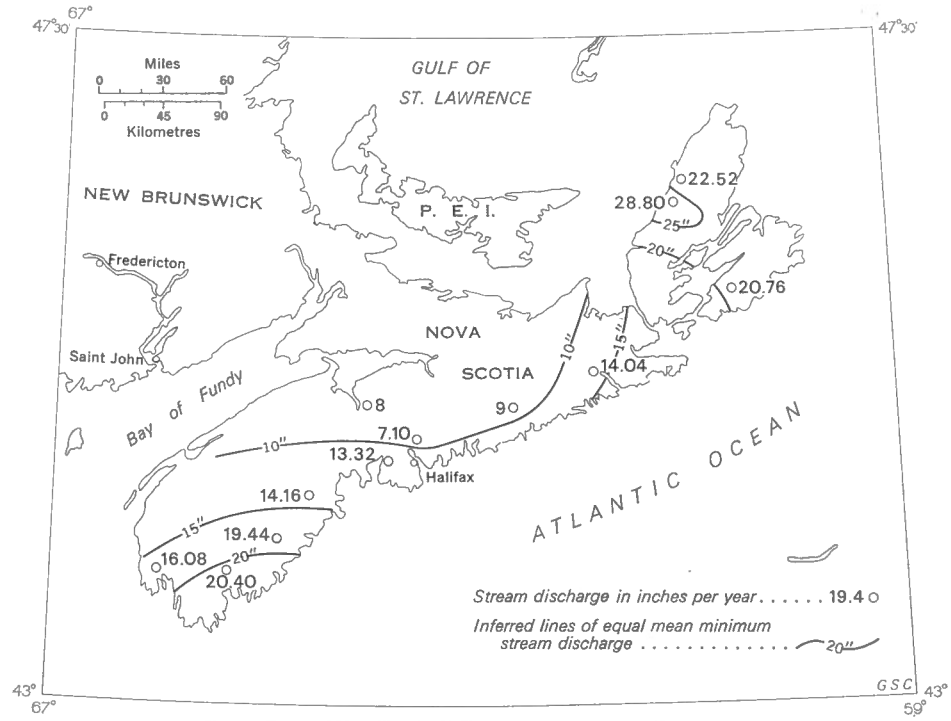


FIGURE 37. Mean minimum stream discharge in Nova Scotia.

river waters are very soft and low in mineralization and are thus very similar in quality to the rivers along the east coast of the mainland of Nova Scotia.

Surface waters of the region have a very low mineral content and are thus soft. Mineralization of the waters is due mostly to the hardness salts, principally to the bicarbonates of calcium and magnesium. Although many rivers are highly coloured, few carry appreciable amounts of suspended matter. Tidal rivers and those flowing through heavily cultivated areas may at times be excessively turbid.

Because of their very low mineral content, high carbon dioxide and low pH, the surface waters are very corrosive. Because the rivers are tidal for most of their courses, the relative content of sodium chloride is often high, especially in waters near the coast.

Groundwater

In Prince Edward Island the groundwater contains calcium bicarbonate water in proportions ranging from 100 to 300 ppm of tds in drilled wells and of about 60 to 100 ppm in springs (Brandon, 1966). Thomas (1960) sampled streams, springs and municipal wells throughout the Atlantic Provinces and found that the water was mainly a calcium bicarbonate type that rarely had a hardness greater than 180 ppm (Fig. 38, *in pocket*).

Apart from local isolated occurrences of high nitrate content, groundwater presents three main chemical problems. These are the following: the high iron, manganese, and sulphate content derived from pyrite in some of the pre-Carboniferous

rocks, especially in Nova Scotia; the high calcium sulphate and sodium chloride content derived from evaporitic rocks, especially from those of the Windsor Group; and sea-water intrusion into coastal areas where groundwater is being overpumped.

Iron

Brandon (*in press*) reports a varying groundwater iron content in the Meguma rocks of Nova Scotia that goes as high as 14 ppm.

Scott (1961) reported highly acid surface water in part of the Shubenacadie drainage system of Nova Scotia. The water had a pH of 2.8 and a total dissolved solids of 2,384 ppm, of which sulphate was 1,497, total iron 158, and manganese 14.8. These constituents were dissolved from pyrite disseminated throughout the Lower Cambrian slates. Small pits resulting from the complete oxidation and removal of pyrite are present in numerous outcrops throughout this area. Also, where the fracture planes are exposed to weathering, the rock has become coated with iron oxide.

Salt and Gypsum

Groundwater containing high total dissolved solids made up mainly of calcium, sodium, sulphate, and chloride occur in evaporitic rocks. These marine rocks comprise a part of the Albert Formation, the Windsor Group and the Codroy Group, all of Mississippian age. A few isolated springs that are high in total dissolved solids occur in the non-marine Pennsylvanian rocks (Hamilton, 1961).

Unsuitable groundwater was first detected in the form of springs. In the last century, in New Brunswick, brine was produced from salt springs by evaporation. The location of the saline springs is shown in Figure 38 (*in pocket*).

New Brunswick

Salt is known to occur in the Albert Formation in the Weldon-Gautreau basin and in the Windsor Group at Dorchester as a plug and along the Anagance axis in the Sussex area. Gypsum and anhydrite occur throughout the Windsor Group.

Saline springs occur in the Sussex area, near Plaster Rock, and in central New Brunswick. The Cornhill, Manhurst, and Gordon Brook saline springs and the brine from the oil wells in the Weldon-Gautreau basin all probably represent salt solutions derived from the Albert Formation (Hamilton, 1961). Brines probably originating from the Windsor Group occur at Plumweseep, Salina, and Plaster Rock. The hypothetical combinations of the constituents of some of these springs are listed in Table VII.

Hamilton (1961, p. 19) noted that the springs at Cornhill, Manhurst, and Gordon Brook were situated near faults. He was unable, however, to determine whether the salt accumulated as plugs or domes as a result of faulting or whether the faults served only as planes along which circulating waters migrated carrying salt from bedded deposits. It is believed that at Plaster Rock the salt is disseminated throughout the gypsiferous shale of the Windsor Group. Hamilton (1961, pp. 15 and 23) believes that at Plumweseep and Salina the salt was concentrated by minor faulting.

Table VII

Chemical Constituents of Some Saline Springs

Hypothetical Combination	Albert Formation, N.B. Cornhill	Windsor Group, N.B.		Codroy Group, Nfld. Fishells Brook
		Plumweseep	Salina	
	ppm	%	%	%
$Ca(HCO_3)_2$	181			
$CaSO_4$	3,604	8.67	18.56	25.85
$MgSO_4$	350			
Na_2SO_4	2,803			
NaCl	32,650	85.46	77.45	50.57
KCl		2.36	1.20	
MgCl			1.14	
$CaCl_2$		0.50		13.84
$MgCl_2$		0.43		8.73

Nova Scotia

Deposits of limestone, gypsum, rock salt, and anhydrite occur in the Windsor Group throughout Nova Scotia. Rock salt 1,500 feet thick occurs at depths of 85 to 3,500 feet below the surface at Malagash, Pugwash, and Nappan, all in Cumberland county. At Mabou, an 86-foot salt-bearing zone was intersected at a depth of 2,260 feet, and at Port Hood salt beds interbedded with anhydrite and shale were noted between 6,090 and 6,625 feet (Collings, 1964, p. 10).

Brine springs occur throughout the whole northern part of Nova Scotia, where the Windsor Group is present (Fig. 38). Cole (1930, pp. 8 and 9) sampled 31 springs; the hypothetical combination of the analyses ranged from 89.74 to 99.21 per cent sodium chloride and 0.30 to 35.97 per cent calcium sulphate. The geology in the vicinity of these springs has been discussed by Hayes (1931).

Newfoundland

Gypsum up to 300 feet thick occurs in the Codroy Group of Mississippian age in the St. George's Bay area of western Newfoundland (Hayes and Johnson, 1938). There are at least 35 localities where gypsum crops out in this area. However, no salt occurrences have been reported, although there are some saline springs.

Four springs relatively high in total dissolved solids were reported by Hayes and Johnson (1938, p. 30). The springs on Fishells Brook and near St. Fintan's are high in sodium chloride (Fig. 38 and Table VII), whereas the spring at Stinking Cove reeks of hydrogen sulphide and the spring near Flat Bay is sulphurous.

Conclusions

In the Appalachian area, the springs high in total dissolved solids consist mainly of sodium chloride with lesser amounts of calcium sulphate. Exceptions occur where springs are in direct contact with gypsum. The geological control of these springs is

often related to faults and hence is complex. The springs can be used as a starting point to explore for salt or gypsum or both. The groundwater flow system should be ascertained and related to possible faults. Then areas of salt-gypsum potential can be delineated and further outlined with gravity surveys. Final proof of the existence of salt and gypsum or of either will come only with drilling, which should readily outline favourable areas and minimize exploration drilling.

Sea-water Intrusion

Sea-water intrusion is significant in Prince Edward Island. To determine its extent, Brandon (1966) calculated the chloride bicarbonate ratio (in equivalents per million) of several analyses. The normal chloride bicarbonate ratio of groundwater on the Island is less than 0.5.

Table VIII lists eight wells that have some sea-water intrusion. In the wells at Miscouche and on Boughton Island, both of which are near the shore and are shallow, sea-water intrusion is slight. The Boughton Island well is below high-tide level and flows at low tide.

Table VIII

Wells Contaminated with Sea Water on Prince Edward Island

Well Location	Chloride (epm)	Carbonate and Bicarbonate (epm)	Ratio	Degree of Contamination
Miscouche (lot 17)	1.25	2.08	0.6	Slight
Boughton Island	1.19	1.80	0.6	Slight
East Point lighthouse	2.47	2.25	1.1	Moderate
Summerside (city well No. 1)	4.18	2.22	1.9	Moderate
Souris wharf	4.47	1.93	2.3	Moderate
Souris lighthouse	4.69	2.01	2.3	Moderate
Federal Building, Charlottetown	15.48	3.02	5.1	Serious
Eastern Fisheries–Souris	19.25	2.10	9.1	Serious

Wells near the shore at Summerside and Souris may have moderate sea-water intrusion. If heavier pumping is undertaken at these places, the salt-water intrusion will probably be a serious problem. The two wells with serious salt-water contamination are deep and are not used for drinking. The salt content of the Charlottetown well was reported to increase with depth.

It is probable that any Island well drilled near the shore to a depth of more than 300 feet will become saline after heavy pumping and that any well drilled inland to a depth of more than 800 feet will also be saline.

Sea-water intrusion also occurs at Shippegan, New Brunswick. There, six fisheries situated adjacent to the sea are pumping water out of the aquifer faster than it is being recharged. The total dissolved solids of these wells, which is essentially sodium

chloride, has been measured by a hydrometer (Crandall, 1964) and is shown in Figure 39 along with the well depth. Salt-water intrusion also occurs in a few other places.

The main cause of salt-water intrusion is overpumping. The estuaries, however, being subject to tidal influences, provide a means for the salt water to penetrate far inland. If overpumping occurs in areas adjacent to estuaries, the normal seaward gradient of the water-table will be reversed and contamination will occur.

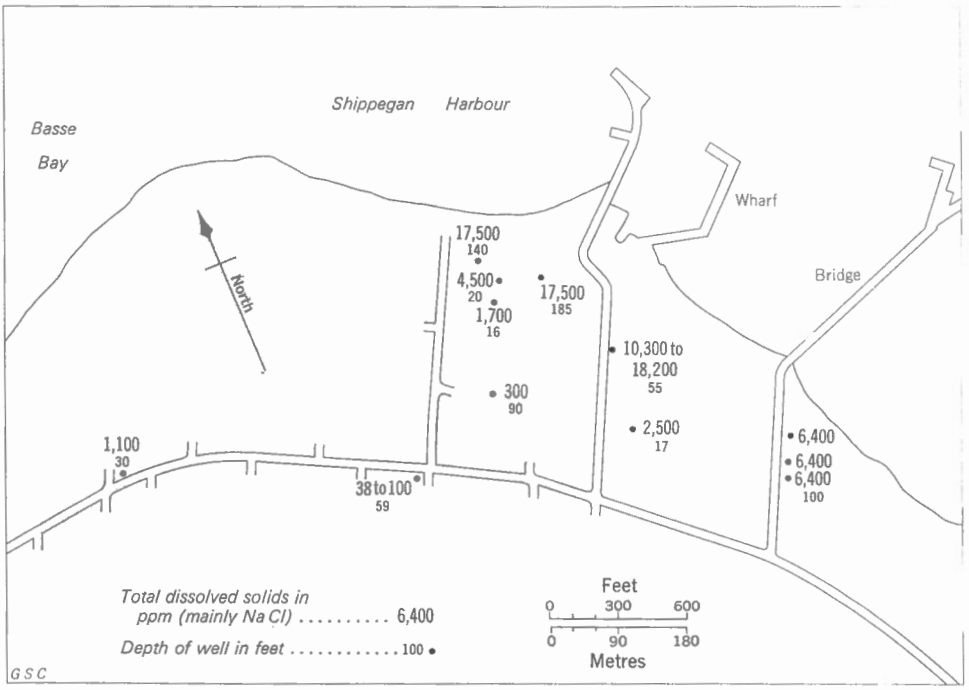


FIGURE 39. Saline wells at Shippegan, New Brunswick, December, 1964.

Flow Systems

Groundwater flow systems have not been studied quantitatively in this region, and only a few qualitative observations have been noted. Although specific information is lacking, the general flow pattern can be ascertained from the physiographic map (Fig. 28), which shows the highlands, uplands, and lowlands.

Surface Flow

All the surface-drainage basins are small compared with others in Canada. The largest, the Saint John, drains about 13,700 square miles in Canada and 7,000 square miles in northern Maine. In much of the area the rivers are short and tidal for a considerable distance upstream. Most of the rivers are controlled by the physiography, but a few like the Saint John have breached upland areas from lowlands.

Groundwater Flow

Groundwater flow is controlled by topography and hence follows the surface drainage systems. This is especially true of the shallow groundwater systems. On a regional scale, in highlands, which are generally considered recharge areas, the depth to the water-table is relatively great and the groundwater flows away from the water-table; in lowlands, which are generally considered discharge areas, the water-table is at or above the land surface and the groundwater flows towards the water-table. Hence groundwater is easier to obtain in lowlands than in highlands. In the Appalachian regions the lowlands comprise the Gulf of St. Lawrence, the plain of New Brunswick, Prince Edward Island and Nova Scotia, and the St. George's Lowland and the Central Lowland of Newfoundland (Fig. 28).

Groundwater recharge will occur everywhere except in discharge areas where the fluid potential of the groundwater is greater than the fluid potential of the surface water. In the Appalachian region, this condition occurs locally in discharge areas that are confined to narrow valleys. Here there is a concentration of flow lines in an area relatively smaller than recharge areas comprising widespread uplands and highlands. Thus, although the amount of groundwater in recharge areas must equal that in discharge areas over a long period, the amounts of groundwater available in discharge areas are relatively larger, for such areas are generally much less extensive.

The amount of groundwater transmitted by a flow system is partly controlled by the permeability of the rock through which it moves. In the uplands and highlands of the Appalachians the rocks are mainly igneous and metamorphic, having fracture permeability that is poorly developed. Thus these rocks transmit little groundwater. On the other hand, in the lowlands (excluding the Notre Dame and Baie D'Espoir Lowlands of Newfoundland, which are filled mainly with igneous and metamorphic rocks of pre-Carboniferous age) the rocks are mainly sandstones, conglomerates, and claystones. These rocks have both fracture and intergranular permeability, which together enable them to transmit far more groundwater than the rocks of the uplands or highland.

Thus, because of their favourable topographic position and the higher permeability of the rocks present, these lowlands are regionally the best places to prospect for groundwater in the bedrock. Only in part of them, however, can large quantities of groundwater be obtained. Specifically, only where the rock is well fractured and is situated in a local discharge area (a valley) can there be high-yielding wells.

Chemical Evidence of Flow Systems

Flow systems can be detected by changes in the chemical composition of the groundwater, but the complete Chebotarev (1955) sequence of groundwater metamorphism has not been detected in the Appalachians, as it has been on the Prairies (J. E. Charron, pers. com.). This is probably due to the lack of deep drill-holes. Softening, the first sign of metamorphism, has been detected, however, in the Moncton map-area (Carr, 1961) and in the Fredericton area (Parsons, *in press*). The end product of groundwater metamorphism, saline water, is also present in the Moncton map-area at depths generally greater than 500 feet.

Discharge of Groundwater Flow Systems

Eventually the fresh groundwater discharges into the oceans or ends up as stagnant saline occurring in a natural trap. Wells situated on the ends of piers penetrating the bedrock often yield fresh groundwater, indicating that fresh groundwater is discharging upward or laterally into the ocean. The water levels within these wells often fluctuate with the tides.

The town wells at Shediac are situated within half a mile of the sea and are pumped at high rates (50, 200 and 400 gpm); yet they show no sign of sea-water intrusion. This indicates that a relatively large zone of fresh-water discharge must occur here; underneath this fresh-water zone, however, is salt water (Fig. 40).

The rise and fall of ocean tides prevents the existence of a sharp interface between fresh water discharging into the ocean and the salt water. This motion creates a zone of diffusion between the two fluids, in which the chloride content varies from that of fresh water, less than 50 ppm, to that of sea water, 19,000 ppm. As the salt water circulates inland from the floor of the sea through the lower part of the aquifer, it becomes progressively diluted with fresh water. As the density of this salt water decreases, the salt water rises, until finally there is no horizontal component of flow and it is carried back to the sea by the flow of the fresh-water system (Fig. 40). The cyclic flow of salt water that takes place during intrusion acts to limit the extent to which salt water occupies the aquifer, because part of the inland flow is continuously returned to the sea (Cooper, *and others*, 1964).

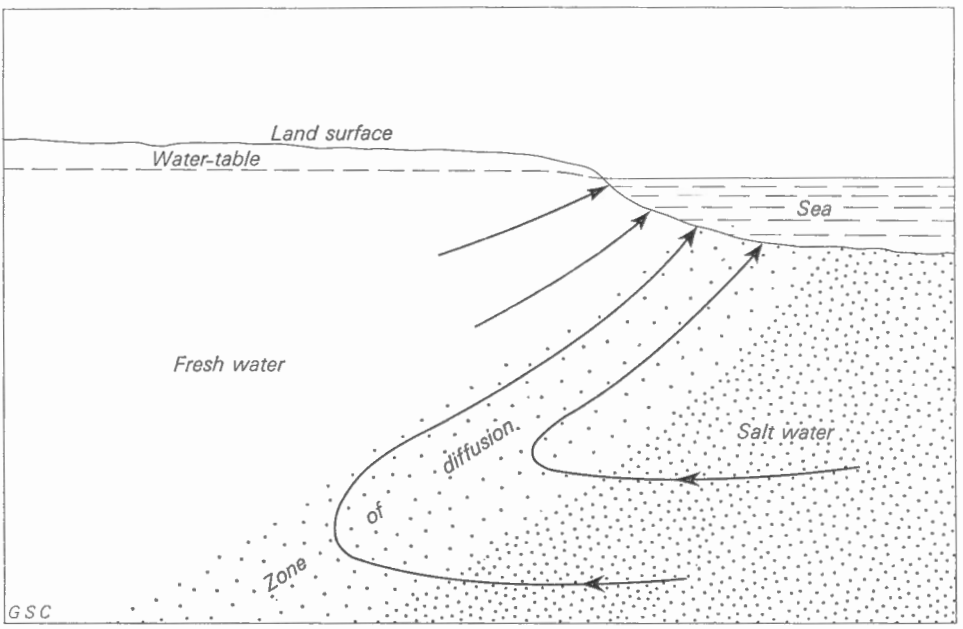


FIGURE 40. Circulation of salt and fresh water adjacent to the sea under normal head conditions (modified from Cooper, *et al.*, 1964).

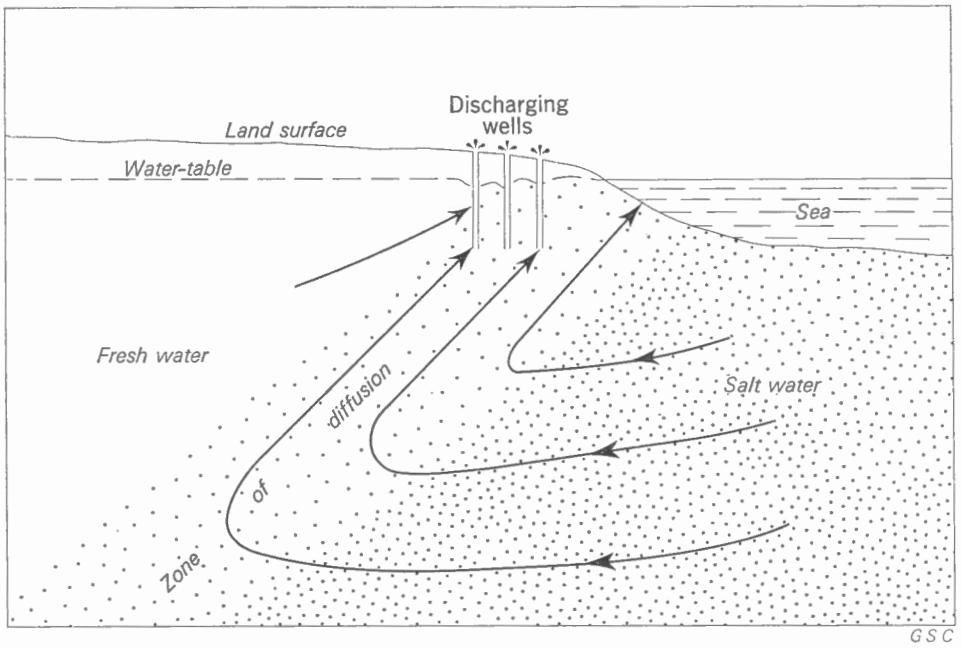


FIGURE 41. Circulation of salt and fresh water adjacent to the sea in an area of heavy pumping.

If there has been heavy pumping adjacent to the sea, the normal groundwater gradient may be reversed, with the result that more extensive sea-water intrusion will occur. Figure 41 shows the fresh water and the salt water flowing towards the discharging wells. The zone of diffusion will peak upward towards these wells. This is what has happened at Shippegan, New Brunswick.

Water Balance

Brandon (1966) is the only one who has attempted to make a water balance for the Maritimes. He made this balance for part of Prince Edward Island, using stream-gauging data from the Dunk, Morell, and Mill Rivers.

The water balance for any simple drainage basin can be expressed as

$$P=E+Rs+Rg.$$

The symbol P represents the inflow of precipitation in inches, and E is the actual evaporation by transpiration from plants and by evaporation from open water and soil moisture. Rs is the measured run-off at a gauging point in a stream and, because groundwater is effluent to the streams here, it combines the surface run-off and the groundwater contribution to the stream above the gauging station. Rg is the groundwater that is discharged from the basin into estuaries or directly along the coast-line into the sea.

The mean annual precipitation recorded at the Experimental Farm at Charlottetown is 42.3 inches, with a maximum of 54.6 and a minimum of 29.8 inches. The

records show that for varying periods there have been successive years of relatively dry weather followed by a succession of wet years. Between 1939 and 1945 there was a period of wet years, which accumulated a surplus of more than 28 inches above the mean.

Evaporation measurements at Charlottetown have already been discussed under *Climate*. For the summer months of the year 1961–62 evaporation, as measured by a black Bellani plate, was 9.2 inches. For the remainder of the year the potential evaporation was calculated by the Thornthwaite method to be about 4 inches.

Run-off comprises storm run-off after heavy rainfall and the spring melt of snow and ice, water that has entered the soil and drained downslope to the rivers, and water that has percolated through the ground and reached the water-table. From here it passes through a slow-moving flow system in the bedrock to a discharge point in the stream or to the sea.

The run-off in inches per square mile of the three river basins for the year 1961–62 is as follows:

Table IX

Run-off of Three Basins in Prince Edward Island, 1961-62

(in inches per square mile)

Basin	Run-off
Dunk	29.5
Morell	27.1
Mill	27.4

Thus it is possible to indicate the order of magnitude of the four factors of the water-balance equation for the wet year 1961–62. These values are approximate and are recorded to the nearest inch on the basis of the precipitation at Charlottetown and an estimated evaporation for the remaining months and the average run-off for the Dunk, Morell, and Mill Rivers.

Table X

Water Balance Estimate for Three Basins in Prince Edward Island

(in inches)

Year	P	≈	E	+Rs	+Rg
1961-62	49	≈	9+4	+28	+8

It is apparent that much more detailed measuring is required over a period of years to produce proper estimates of the water balance.

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Chapter IV

ST. LAWRENCE LOWLANDS HYDROGEOLOGICAL REGION

by J. S. Scott

Introduction

The St. Lawrence Lowlands occupy an area of about 43,000 square miles lying south of the Canadian Shield and extending northeasterly, although not continuously, from the northwest end of Lake Huron and the west end of Lake Erie to Anticosti Island. The area has developed into the most important industrial and agricultural region in the country and contains about half the total population of Canada. As a consequence, the withdrawal of groundwater in the region by municipalities with a population greater than 1,000 (Can. Mun. Utilities Ann. Directory) is 73 per cent of the total groundwater withdrawal by such municipalities throughout Canada. It is probable that rural and industrial users of groundwater in the region obtain a similar proportion of the national volume obtained by such users.

Physical Features

The St. Lawrence Lowlands, a plains-like area underlain by unfolded Palaeozoic rocks, may be divided into three parts. The western and central parts are separated by the Frontenac axis, a projection of the Canadian Shield that crosses the St. Lawrence River between Kingston and Brockville. A similar projection of the Canadian Shield, the Beauharnois axis, crosses the central part of the Lowlands at the confluence of the Ottawa and St. Lawrence Rivers. Although the Beauharnois axis does not create a major physiographical break it is the locus of the oldest rocks within the central part of the Lowlands and separates the Palaeozoic rocks of the Ottawa area from those of the St. Lawrence Valley. Anticosti Island, Mingan Islands, and a narrow strip of the north shore of the St. Lawrence River form the eastern part, which is separated from the central part of the Lowlands by about 360 miles of the St. Lawrence River.

West of the Frontenac axis, the physiographical region is separated into two parts by the Niagara Escarpment, a prominent physiographical feature extending west

from the Niagara River to Hamilton, thence northwesterly to Georgian Bay, along the Bruce Peninsula to Manitoulin Island. In the Niagara Peninsula, the escarpment rises abruptly 250 to 300 feet above lower ground to the northeast and attains an elevation of about 650 feet. South of Georgian Bay, it forms part of a narrow belt of upland where elevations between 1,700 and 1,800 feet are reached as at Collingwood and Orangeville. These elevations are about 750 feet above the lowlands that extend southeastward from the cuesta scarp towards Lake Ontario.

From Lake Ontario, at an elevation of 246 feet, the land east of the escarpment rises gently northward to a maximum height of more than 1,000 feet in the area west of Lake Simcoe. At a few places east of Lake Simcoe, a second escarpment with an abrupt drop to the Precambrian rocks of the Canadian Shield marks the north boundary of the St. Lawrence Lowlands; however, the north boundary is more typically transitional from the lowlands to the Shield. West from the escarpment towards Lake Huron at an elevation of 580 feet and southwest toward Lake Erie at an elevation of 572 feet, land elevations decrease gradually through an area of rolling topography with low relief.

The central part of the St. Lawrence Lowlands occupies the area between the Ottawa and St. Lawrence Rivers, and extends from the east side of the Frontenac axis along the St. Lawrence River valley to a point a few miles east of Quebec City. The north boundary of the central part is marked by overlap of Palaeozoic formations of the lowlands with the Precambrian crystalline rocks of the Canadian Shield. East of the St. Lawrence River, a fault zone known as "Logan's Line" separates the lowlands from the hilly Appalachian region of folded rocks. Elevations within the central part of the lowlands do not exceed 500 feet above sea-level except for a few isolated hills of intrusive igneous rocks that project through the surrounding Palaeozoic formations. East of Montreal, the land rises from elevations of 100 feet at the borders of the St. Lawrence River to heights of about 300 feet at the boundaries of the Lowlands.

Anticosti Island, in the Gulf of St. Lawrence, is the largest section of the eastern part of the St. Lawrence Lowlands. The Mingan Islands in Jacques Cartier Passage and a narrow strip of the adjacent mainland are also part of the Lowlands hydrogeological region, but because of their relative insignificance from the standpoint of groundwater this area is not included in this report. Anticosti Island is about 125 miles long and its greatest width, near the middle of the island, is about 35 miles. It has the form of a cuesta with a north-facing escarpment, and consequently the highest elevations, about 600 feet, are nearer the north coast of the island. Terraces are a notable physiographical feature of the island, occurring on both the north and south sides. More than twenty terraces are known with the highest at an elevation of 400 feet above sea-level. Along the scarp face of the north shore the terraces are narrow and a terrace may in places merge with the ones above and below, whereas on the south shore they may be several miles wide. Although most of the terraces are covered with gravel, those above the 250-foot level are the result of bedrock control whereas those below have formed mainly as a result of marine erosion.

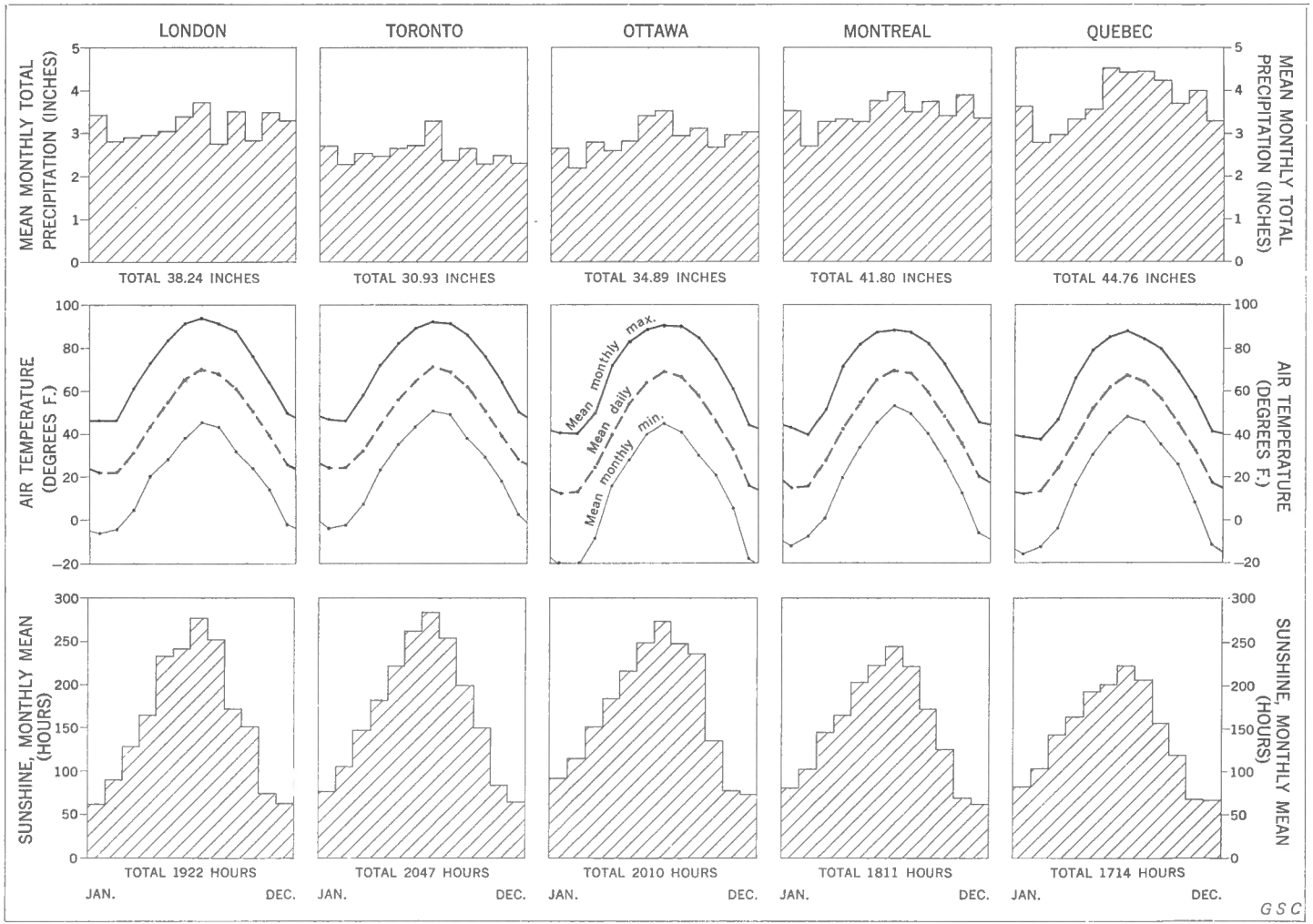


FIGURE 42. Climatic data for stations in the St. Lawrence Lowlands hydrogeological region.

Climate

The hydrogeological characteristics of the St. Lawrence Lowlands area are determined largely by the prevailing climatic and geological conditions. The interrelationship of such factors as precipitation, temperature, wind velocity, and sunshine with the hydrological characteristics of the geological materials determine the proportion of the water balance that is available for groundwater recharge. The western and central parts of the Lowlands are contained within the southeastern climatic region whereas the eastern part of the Lowlands lies within the northern or subarctic climatic region (Boughner and Thomas, 1960).

The southwestern tip of Ontario extends farther south than any other part of Canada, which fact, along with the moderating influence of the Great Lakes, gives peninsular Ontario a much milder climate than that of the northern areas of the country. One of the major storm tracks of the continent passes through the lower lakes region producing wide variations in day to day weather, particularly in winter. Thus, periods of extreme conditions of severe cold or excessive heat are not prolonged. To the north and east, the Lowlands are subjected to the same climatic influences as the lower lakes region of Ontario, but this area is without the protection afforded by the Great Lakes. Maritime influences, however, serve to moderate the climate of Anticosti Island as compared with continental areas to the north.

Precipitation throughout the Lowlands is generally uniformly distributed during the year except in the eastern part of the St. Lawrence Valley where a tendency towards maximum precipitation in the summer and early autumn exists. The mean total precipitation of 38 inches at London, Ontario, decreases to 31 at Toronto, and increases northeastward throughout the Ottawa–St. Lawrence Valleys to almost 45 inches at Quebec City (Fig. 42).

The mildest winter temperatures are recorded around Lakes St. Clair and Erie and in the Niagara Peninsula where January temperatures average 24°F reflecting the influence of the Great Lakes. Considerably colder winters occur to the north and east as evidenced by the January mean temperature of 12°F at Ottawa and Quebec City and 15°F at Montreal. Mean daily temperatures attain a maximum of about 70°F in July throughout the Lowlands area. Total mean monthly hours of bright sunshine are at a maximum at Toronto and decrease towards the north and east. The total at London is also slightly less than the total at Toronto.

A qualitative assessment of the effect of climatic conditions on fluctuations of groundwater levels can be obtained from a composite hydrograph from a well in a shallow aquifer (Fig. 43). Observation well No. 15 of the Ontario Water Resources Commission is located within the Thames River watershed in the city of London. The well is 2 inches in diameter drilled to a depth of 40 feet into a sand and gravel aquifer. Nine years of monthly observations (1948–56 incl.) were plotted to form the composite hydrograph shown on Figure 43. The well does not appear to be unduly affected by local pumping during the period of the records; thus changes in water level are due mainly to climatic changes. The composite hydrograph clearly shows an increase in water level from a winter low at the end of January to a seasonal high at the end of March. A decrease in water level then takes place until the end of

August at which time the water level begins to increase to a secondary maximum at the end of December.

It is apparent from the relatively uniform distribution of precipitation (Fig. 42) that other factors strongly influence the quantities of precipitation available as recharge to the aquifer. Maximum recharge takes place in the spring at the time of spring break-up and before increases in evaporation by solar energy and transpiration by plants occur. Rapid increases in solar energy take place at the end of March and continue through to the end of July as shown by temperature and bright sunshine data on Figure 42. During this period most of the precipitation is consumed by plants, runs off directly, or is returned to the atmosphere by evaporation, and the amount of groundwater stored by the aquifer is decreased by gravity drainage to areas of natural discharge. Groundwater recharge again occurs during the fall consequent upon a decrease in solar energy and demands by vegetation. The recharge process apparently continues throughout the early winter until the infiltration capacity of the recharge areas is reduced by freezing of water in the pore spaces.

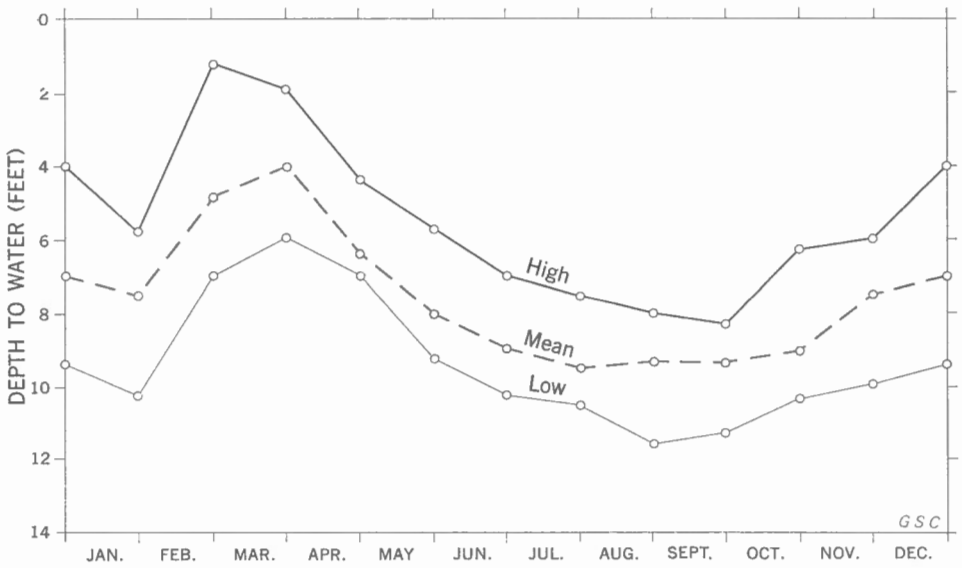


FIGURE 43. Composite hydrograph Ontario Water Resources Commission observation well No. 15, 1948-56, London, Ontario.

Variations in the magnitude of the climatic factors occur from year to year and these variations are reflected by groundwater levels. In the composite hydrograph the difference between the highest and lowest level recorded for any given date varies between 3 and 6 feet. The annual trend of the hydrograph, however, remains the same; variations take place only in the magnitude of the water-level fluctuations. A similar annual trend of water-level fluctuations could be expected in shallow aquifers to the north and east throughout the Lowlands area because of the general similarity of climatic conditions.

Hydrogeology

Throughout the St. Lawrence Lowlands, bedrock formations are of Palaeozoic age with the exception of older rocks of Precambrian age at the mouth of the Ottawa River and the small masses of younger intrusive rocks of Cretaceous age in the Montreal area. The entire Lowlands area was subjected to glaciation during the Pleistocene epoch; thus the bedrock is covered, in most places, by unconsolidated sediments of glacial and related origin.

The water-bearing characteristics of the geological materials are dependent upon the extent and degree of development of openings within the materials as is discussed in detail in Chapters I and II. The most important openings that permit movement of groundwater are fractures in the bedrock formations and intergranular pores in the unconsolidated sediments. Significant variations in the lithology of the geological materials occur both stratigraphically and areally throughout the Lowlands area. As a consequence, the aquifer characteristics of the geological materials vary from place to place throughout the area.

Sufficient detailed studies have not been made, as yet, to permit a detailed quantitative assessment of the groundwater potential of the Lowlands. Reports of the Ontario Department of Mines, Ontario Water Resources Commission, Quebec Department of Natural Resources, and the Geological Survey of Canada, however, provide valuable basic data from which a preliminary qualitative assessment of the groundwater resources of the Lowlands can be made.

During the Palaeozoic era, the Lowlands area was inundated by seas in which sediments accumulated and the sediments were subsequently indurated to form rock layers. The seas, however, did not at all times extend continuously over the entire region; thus the rock sequence is not the same everywhere and contemporaneous sediments may vary greatly in lithology. Geological description of the Lowlands in published reports has been facilitated by subdividing the region into several parts as follows: Ontario, west of the Frontenac axis; Ontario, east of the Frontenac axis; Quebec; and Anticosti Island. The same subdivisions are used here to describe the hydrogeology of the area.

Bedrock

Ontario, West of the Frontenac Axis

Marine sedimentary strata of Cambrian, Ordovician, Silurian, Devonian, and Mississippian ages underlie this part of the Lowlands region. These rocks rest upon the uneven surface of the Precambrian Shield and dip gently southwestward from the Shield towards the Michigan Basin and southeast into the Appalachian Geosyncline. The structure of the rocks is such that they form successive northwest-trending belts as is clearly shown on Figure 44 (*in pocket*).

Groundwater associated with bedrock formations is mineralized to varying degrees depending upon the solubility of the minerals contained in the rocks and the rate and length of flow of groundwater through the rocks. The oldest Palaeozoic rocks in the area are mainly sandstones of the Potsdam Formation of Cambrian age.

Outcrops of the formation occur only in the Kingston area adjacent to the Frontenac axis as the formation is restricted mainly to the area near the eastern end of Lake Ontario. In the central and eastern parts of the Lowlands, the formation is generally a good source of groundwater but it is not a significant aquifer in the area west of the Frontenac axis.

Rocks ranging in age from Middle to Upper Ordovician underlie the area extending southwest from the Precambrian Shield to the base of the Niagara Escarpment. The oldest rocks are soft, red and green shales with beds of sandstone and arkose that have been termed Shadow Lake Formation. The formation, which does not exceed 50 feet in thickness, outcrops in the general area east of Lake Simcoe. Succeeding strata of the Simcoe Group (Black River-Trenton) are dominantly thin- to thick-bedded, grey limestones that range in thickness from 550 to 1,000 feet. The development of both vertical and horizontal joints has created a moderate aquifer potential for rocks of the Simcoe Group. Wells generally yield an average of less than 10 gpm (Fig. 44). In some areas underlain by these rocks, however, well yields are less than 1 gpm and groundwater supplies are obtained from aquifers in the overlying surficial deposits.

Black, petroliferous, calcareous shales and soft, brown and blue shale of the Whitby Formation are in sharp contact with the underlying rocks of the Simcoe Group. The formation ranges from 80 to 290 feet in thickness, and the rocks generally occur with fissile shale partings or as thin beds. Although jointing is well developed in some of the rocks, the aquifer potential of the formation as a whole is poor as is the quality of water derived from these rocks. In the area north of Whitby, wells yield 5 gpm or less but, in general, throughout the area underlain by these rocks, groundwater supplies are obtained from aquifers in surficial materials.

The uppermost beds of the Whitby Formation are in sharp contact with the overlying rocks of the Georgian Bay Formation, which are an alternating sequence of shale and limestone with an aggregate thickness that varies between 200 and 600 feet. The upper member of the Georgian Bay Formation is limestone and attains its greatest thickness on Manitoulin Island. Although the rocks are thin to medium bedded and well jointed, the formation is not a good aquifer as well yields are only a few gallons per minute and the water commonly is highly mineralized. Throughout most of the area underlain by the Georgian Bay Formation groundwater supplies are obtained from aquifers in the overlying surficial deposits.

The youngest Ordovician strata are red, red mottled green, and green shales of the Queenston Formation which are in sharp contact with the underlying rocks of the Georgian Bay Formation. In the Georgian Bay area, grey organic limestone is interbedded with the shales of the Queenston Formation, which shows a considerable variation in thickness from place to place with a general thickness of 240 feet. At Niagara Falls the Queenston Formation attains a thickness of 1,200 feet. Colour banding, a common feature of the shales of the Queenston Formation, transects the poorly developed bedding of the shales. As seen in outcrop the shales do not display a prominent joint pattern; however, the high degree of orientation of the clay minerals within the shale impacts a blocky fracture to these rocks.

The average yield of wells that obtain groundwater supplies from the Queenston Formation is 5 gpm although substantially larger quantities are obtained in some localities. In the Niagara Peninsula, where the formation is dominantly shale, the aquifer potential is poor as wells generally yield but small quantities of highly mineralized water. Aquifer conditions improve towards Georgian Bay which may be due to the presence of limestone in the upper part of the formation in that area.

At the end of the Ordovician period marine sedimentation ceased as the seas withdrew. An interval of erosion of the previously deposited sediments ensued prior to a renewed invasion by the seas in which rocks of the Silurian system were deposited.

Rocks of Lower and Middle Silurian age constitute the Clinton and Cataract Groups. Formations within the groups contain a diversity of rock types such as white and grey sandstone, grey and red shale with minor gypsum, and dolomite. The rock groups, which have a total thickness of 190 to 230 feet, are not utilized as aquifers to any great extent. In most places they are overlain by younger strata with greater aquifer potential. However, in those localities along the Niagara Escarpment close to the outcrop area of these rocks, highly mineralized groundwater supplies are obtained from wells yielding 2 or 3 gpm. Shales of low permeability within the Clinton Group restrict the downward movement of groundwater. As a consequence, many springs occur along the face of the Niagara Escarpment at the contact of the shales with the overlying fractured dolomite.

A sequence of Middle Silurian strata consisting of the Lockport, Amabel, and Guelph Formations cap the Niagara Escarpment. Rocks of these formations are mainly bluish grey and grey (Lockport–Amabel) and brown (Guelph), very fine to finely crystalline dolomites containing organic reef structures. Bedding characteristics of the rocks vary from thin to thick, and both vertical and horizontal joints are well developed. The aggregate thickness of these formations ranges from 150 to 650 feet. In general, rocks of these formations constitute some of the better aquifers in the western part of the Lowlands. Average well yields are of the order of 10 gpm but in some places wells yield in excess of 600 gpm. The amount of mineralization of waters derived from these rocks is generally not excessive except where small inclusions of gypsum or disseminations of sulphide minerals have locally produced highly mineralized waters.

The dolomites of the Middle Silurian formations grade upward into the Salina Formation that comprises brown dolomite, grey limy shale, gypsum, anhydrite, and salt. The formation ranges in thickness from 300 to more than 1,500 feet with the greatest thicknesses in the southwestern part of Ontario. Salt beds within the formation are confined to the area west of London where they occur at depths of 1,000 feet or more and thus are beyond the range of water wells in this area. Bedding characteristics of the formation range from thin to medium for the dolomites and shales to thick for the evaporite beds. Details of the fracture characteristics of the rocks are not known but it is probable that both vertical and horizontal jointing have been developed in these rocks. Wells penetrating the upper part of the formation generally yield from 5 to 15 gpm. The overall value of the formation as an aquifer, however, is somewhat diminished by the highly mineralized waters that are commonly encountered.

The youngest Silurian strata, which conformably overlie the Salina Formation are cream and tan dolomites with oölitic zones that have been assigned to the Bass Island Formation. These rocks occur in medium to thick beds with an aggregate thickness of 60 to 395 feet. Jointing is well developed and consequently the formation is a good aquifer from which average well yields of 5 to 15 gpm are obtained.

At the close of the Silurian period uplift of the region caused a recession of the seas and the onset of an interval of erosion. Lower Devonian time began with the return of the sea and the initiation of sedimentation of the Devonian system of rocks.

The oldest Devonian rocks are light grey, medium-grained sandstones of the Oriskany Formation that rest upon the eroded surface of the Silurian dolomites. These sandstones are known only in the Niagara Peninsula where they attain a thickness of as much as 20 feet. The limited thickness and distribution of the formation is due, in part, to an interval of erosion that occurred prior to deposition of rocks of the overlying Bois Blanc Formation. The Oriskany Formation is, therefore, only of minor significance as an aquifer.

Rocks of the Bois Blanc Formation are brown dolomitic limestone with abundant chert and minor sandstone. The formation ranges from 65 to 265 feet in thickness and disconformably overlies the Oriskany Formation or, where the Oriskany is absent, Silurian strata. A good aquifer potential has been developed in the formation by the presence of a well-developed joint system superimposed upon the thin- to medium-bedding characteristics. Average well yields range from 10 to 30 gpm of moderately mineralized water.

Bois Blanc strata are overlain by a sequence of dominantly carbonate rocks of the Detroit River Formation. These younger rocks comprise brown fine crystalline limestone, minor sandstone, granular dolomite, and beds of lithographic limestone with very thin bituminous partings. Organic reef structures, which may be cavernous in places, are found in these rocks. Detroit River strata occur in medium to thick beds except for massive reef structures and the rocks generally display a well-developed system of joints. Groundwater supplies from the formation are generally good with well yields averaging 10 to 30 gpm. The quality of water from shallow depths in the formation is usually good with a tendency to become more highly mineralized at greater depths.

Brown and grey limestone and dolomite (with minor nodular chert and sandy limestone) are the principal rock types contained within the Norfolk Formation that overlies Detroit River strata. The formation attains a thickness of about 250 feet, and the rocks, which occur in medium to thick beds, are transected by a system of well-developed vertical joints. Average well yields range from 1 to 10 gpm throughout most of the area underlain by the Norfolk Formation; thus these rocks form a fairly good aquifer. In the area adjacent to Lake Erie, however, the formation is not exploited as an aquifer as groundwater supplies are obtained at shallower depths in the overlying surficial deposits.

A sequence of two alternations of grey shale and argillaceous limestone constitute the Hamilton Formation which conformably overlies the Norfolk Formation. These younger rocks range in thickness from 140 to 300 feet. Strata within the formation

occur in thin to medium beds and the rocks are probably jointed but the extent of joint development is not known. In comparison with the older strata the Hamilton Formation is only fair as an aquifer, as average well yields range from 1 to 5 gpm. Throughout most of the area underlain by the formation groundwater supplies are obtained from aquifers in the overlying surficial deposits.

The youngest Devonian rocks in the area are brown and black, bituminous fissile shales of the Kettle Point Formation. The shales commonly contain disseminations of pyrite and marcasite, and at Kettle Point on Lake Huron, where the formation is typically exposed, calcareous concretions up to 3 feet in diameter (commonly referred to as "kettles") are contained within the rocks. The formation is, in general, a poor aquifer; yet in areas where other sources of groundwater are not available, well yields of a few gpm are obtained from the upper part of the formation.

Grey and black shale and grey sandstone of the Port Lambton Formation overlie the Kettle Point Formation. These rocks are of Mississippian age, the youngest Palaeozoic rocks in Ontario, but they are known only from well cuttings. The aquifer potential of the Port Lambton Formation is not known but it would be of limited significance in Ontario because of the restricted distribution of the formation.

Ontario, East of the Frontenac Axis

In eastern Ontario and immediately adjacent to Quebec, Palaeozoic rocks of Cambrian and Ordovician age are present in a well-defined basin. The basin is bounded on the west by the Frontenac axis, on the north by the Canadian Shield, on the east by the Beauharnois axis, and on the south by the St. Lawrence River although similar physiography and geology extend south to the Adirondack Mountains in the United States. Over the greater part of the basin the rocks are flat lying or gently undulating except where they are tilted within fault blocks or near the principal fault zones. All the faults are steeply dipping and the principal set trends east-southeast. The major effect of faulting of these rocks has been the placement of older rocks in abrupt contact with younger rocks as is shown on Figure 44. Although faulting and associated fracture development in rocks increase the ability of the rocks to transmit water, many of the faults and fault zones in this area are actually barriers to groundwater movement as the fracture zones have subsequently been sealed by fillings of calcite.

The uneven Precambrian surface in the Ottawa area is overlain first by grey and yellow sandstone and siltstone of the Nepean Formation. Medium- to coarse-grained quartz sandstone locally altered to quartzite is found in the basal beds of the formation. Beyond the Ottawa area the formation is typically siltstone and may be known by the term Potsdam. Bedding characteristics are variable from thin to massive and the thickness of the formation also varies from a few inches to about 800 feet, the latter found only in wells as outcrop thicknesses seldom exceed 200 feet. Both vertical and horizontal joints are present in the rocks with the best development of joints in the upper and lower beds of the formation, and consequently these zones form the best aquifers within the formation. Average well yields of 10 gpm are commonly obtained from aquifers in the formation, and in places where fractures are well developed throughout a substantial thickness of the formation, wells may yield several

hundred gpm. The total amount of dissolved solids in groundwater obtained from the Nepean Formation is generally excessive, but the amounts vary considerably from well to well depending upon the location within the flow system from which the water is obtained.

A probable disconformity separates the Nepean Formation from the alternating grey sandstone and sandy dolomite of the overlying March Formation. The thickness of the formation is 25 to 30 feet and thus it forms a transitional unit between the Nepean Formation and the rusty dolomite and limestone of the Oxford Formation which has a thickness of about 300 feet. March and Oxford strata occur in thin to thick beds that are transected by a well-developed system of vertical joints. Several aquifer zones exist within the March and Oxford Formations and average well yields vary from 5 to 15 gpm. Substantially larger well yields have been obtained from aquifers in these formations in places where fractures are well developed and conditions for adequate recharge to the aquifers exist.

A disconformity separates the Oxford Formation from the green shale and sandstone lenses of the Rockcliffe Formation which in turn is overlain by impure grey limestone, dolomite, and shale of the St. Martin Formation. Rockcliffe and St. Martin strata occur in thin to medium beds and display well-developed jointing. The aggregate thickness of the two formations varies from 190 to 315 feet. These formations are hydrologically similar to the underlying formations and the aquifer zones within them produce average well yields of 10 gpm.

Throughout this area, the St. Martin Formation is overlain by grey limestone with minor shale partings that constitutes the Ottawa Formation. These rocks attain a thickness of as much as 570 feet. The lower contact of the Ottawa Formation is unconformable in the western part of the basin where these younger rocks rest upon successively older strata. Bedding characteristics within the Ottawa Formation vary from thin to thick with the thickest beds in the middle and upper parts of the formation. Both horizontal and vertical joints are well developed in these strata and thus the rocks are hydrologically similar to the underlying carbonate formations. Well yields vary from 5 to 15 gpm from aquifers in the fractured zones of the rock. In the Ottawa area several hundred feet of the Ottawa Formation are composed of massive limestone that yields no water. Low well yields from the Ottawa Formation are also indicated in the area around Casselman (Fig. 44), but there the presence of adequate aquifers in the overlying surficial deposits reduces the utilization of the bedrock aquifers.

A transition from the limestone of the Ottawa Formation to the shales of the Billings Formation, formed by a sequence of dark grey, fine-grained limestone with interbedded shale, has been assigned the term Eastview Formation. Eastview strata have a maximum thickness of only 20 feet and are of limited areal extent; thus from the standpoint of hydrogeology these rocks are included with the upper part of the Ottawa Formation.

Rocks of the overlying Billings Formation are brown and black, pyritiferous, fissile shale that attains a thickness of from 260 to 300 feet. These dark shales are, in turn, overlain by 500 to 800 feet of grey shale with thin, impure beds of limestone or

dolomite that constitute the Carlsbad Formation. Both the Billings and Carlsbad Formations are of limited areal extent in the area east of Ottawa. The rocks are locally utilized as aquifers for domestic purposes and some wells yield as much as 20 gpm. Average well yields, however, are much less and the water obtained from these formations is commonly highly mineralized.

Grey shale and dolomitic limestone of the Russell Formation are overlain by red shale of the Queenston Formation. The aggregate thickness of these units is about 100 feet. Both formations have a very limited distribution and are found only in the area northwest of Russell, Ontario. The hydrological properties of these formations are not known but they are probably similar to those of the Queenston Formation in the Niagara Peninsula where wells in the formation yield an average of 5 gpm.

Quebec

The St. Lawrence Lowlands area east of the Beauharnois axis is also underlain by Palaeozoic strata of which only the Cambrian and Ordovician systems are represented, and these are intruded locally by Cretaceous igneous rocks as exposed in the Monteregian Hills. Although the rocks are essentially flat lying, they have been warped into an elongate syncline or trough. The oldest rocks are exposed around the periphery of the syncline, adjacent to the Canadian Shield and the Beauharnois axis, and the younger rocks occupy the central part of the syncline. The southeastern boundary of the area is marked by the fault zone (Logan's Line) that forms the northwestern limit of the folded Appalachian region. Some disturbed strata, therefore, can be found in the Lowlands area adjacent to the fault zone. In contrast with the maximum thickness of about 2,200 feet of Palaeozoic rock in eastern Ontario, at least 7,500 feet of these strata are known to be in the central part of the syncline near Nicolet, Quebec.

Few published hydrogeological data are available for the Quebec part of the St. Lawrence Lowlands so it is not possible to show the distribution of average well yields for that area. An indication of the groundwater potential of the area, however, can be obtained from existing hydrogeological reports published by the Quebec Department of Natural Resources and the Geological Survey of Canada and by a comparison of the geology of this area with the geology of the areas to the west.

The oldest rocks within the Quebec part of the St. Lawrence Lowlands are crystalline limestone, granite, and gneiss of Precambrian age that occur along the Beauharnois axis at the western end of the area and within the Canadian Shield along the northern boundary of the Lowlands. Fractures commonly extend a few tens of feet into these rocks; thus local aquifers capable of supplying domestic needs can be found within the Precambrian rocks.

The Precambrian rocks are first overlain by thin- to medium-bedded, white quartz sandstone and conglomerate of the Potsdam Formation, which is, therefore, similar in lithology to the Nepean Formation of eastern Ontario. Rocks of the Potsdam Formation occur in thin to medium beds that are transected by a well-developed joint system. The formation attains its greatest thickness in the southwestern part of the area. The strata thin toward the northeast and are not present in

the Nicolet area. In the area west of Montreal, aquifers in the Potsdam Formation have produced as much as 600,000 gpd; thus the aquifers are suitable for municipal and industrial use. From the standpoint of potential well yield and quality of water, aquifers in the Potsdam Formation are the best of those found in the Palaeozoic rocks.

An interval of marine regression and erosion followed deposition of the Potsdam sediments as indicated by the presence of reworked sands in the basal beds of the overlying Beekmantown Group. Formations within the Beekmantown Group comprise minor sandstone and dolomite in the lower part of the group and grey dolomite with thin shale partings in the upper part. Bedding characteristics of these rocks vary from thin to thick and a joint system has been moderately developed. The thickness of the group varies from a few to more than 1,000 feet.

In the area west of Montreal many high capacity wells utilize aquifers in the fractured dolomites of the Beekmantown Group. Well yields from 100,000 to 700,000 gpd have been encountered but the water is generally hard. It is probable that average well yields of 5 to 15 gpm would be obtained from smaller wells as is the case with aquifers in equivalent strata (March and Oxford Formations) in eastern Ontario.

Middle Ordovician strata of the Chazy Group are separated from Beekmantown strata by a stratigraphic break representing an interval of erosion. Rocks of the Chazy Group are mainly limestone with dolomite, shale, and sandstone layers and have an aggregate thickness ranging from a few feet to 650 feet. A minor disconformity separates rocks of the Chazy Group from the overlying limestones of the Black River and Trenton Groups that range in total thickness from 600 to 2,500 feet.

Carbonate rocks of the Chazy, Black River, and Trenton Groups are characterized by thin to thick beds and both vertical and horizontal jointing are moderately well developed. The extent of fracturing and the development of other openings in these rocks is not everywhere the same; thus there is considerable variation in their aquifer potential. North of Quebec City a well with 170,000 gpd capacity was developed in Trenton limestone, and another well in the same rocks 40 miles west of Quebec City has a yield of 400,000 gpd. Wells yielding from 100,000 to 600,000 gpd have also been developed in aquifers in Black River and Chazy strata. These high capacity wells, however, are exceptions as most wells yield much smaller quantities probably of the order of 5 to 15 gpm, as is produced from aquifers in the stratigraphic equivalents of these groups in eastern Ontario.

Upper Ordovician strata in Quebec are represented by rocks of the Utica, Lorraine, and Richmond Groups. In contrast with the underlying carbonate sequence, rocks within the groups are mainly shale with minor interbeds of limestone and sandstone. The three groups attain an aggregate thickness which may exceed 6,000 feet. Vertical and horizontal fractures are not well developed in the shales but are present to a sufficient extent to permit some movement of groundwater through these rocks. Although the aquifer potential of the shales is generally poor, well yields of a few gallons per minute are obtained for local domestic supplies.

The Palaeozoic sediments in the Montreal area and the area east of Montreal were intruded during Cretaceous time by plug-like masses of alkaline igneous rocks including such varieties as essexite and nepheline syenite. These intrusions form the

Monteregian Hills, some of which rise more than 1,000 feet above the surrounding plains.

In general, the intrusive rocks are impervious except where fractures have been developed within them. Fracturing is most extensively developed in the zone of contact of the intrusive rock with the sedimentary rock. Well yields of from 50,000 to 80,000 gpd have been obtained from aquifers in the fractured zone. The limited areal extent of the intrusive rocks is such that these rocks are not significant aquifers on a regional basis.

Anticosti Island

Anticosti Island is underlain by rocks of both the Ordovician and Silurian Systems with a total thickness of about 2,400 feet. The strata dip gently southwards with an average angle of less than 2 degrees. Two systems of nearly vertical joints have developed in the rocks; the most prominent set trends N50°–60°W and is intersected, almost at right angles, by the secondary joint set. It is probable that horizontal bedding joints have also developed in these rocks.

Ordovician strata are subdivided into three formations: the oldest rocks exposed are limestone and shale of the English Head Formation overlain, in turn, by interbedded limestone and shale of the Vauréal Formation and the Ellis Bay Formation which comprises shale and limestone on the south shore and sandstone and limestone on the north shore.

The succeeding Silurian strata of the Becscie, Gun River, Jupiter, and Chicotte Formations consist of limestone and shale in the lower part of the sequence, with limestone dominant in the upper part.

Water supply for the relatively small population of the island is obtained entirely from surface sources, so that in the absence of a demand for groundwater supplies no water well data exist. The presence of groundwater, however, may be inferred from the presence of perennial streams that must derive a substantial portion of their low flow from groundwater discharge.

Aquifers undoubtedly exist in the fractured Palaeozoic rocks of Anticosti Island and it is probable that wells yielding an average of 5 to 10 gpm could be developed in these rocks as has been done in similar rocks in the central and western parts of the Lowlands.

Preglacial Erosion

Prior to the advent of glaciation, the Palaeozoic sedimentary rocks of the St. Lawrence Lowlands had been subjected to millions of years of erosion by rivers. Major river valleys were established in the more easily eroded rocks such as shales, which left the more resistant dolomites and limestones to form the adjacent uplands that were drained by tributaries connected to the main streams.

In southern Ontario, where the rocks dip southwest from the Canadian Shield into the Michigan Basin, the differential erosion between hard and soft strata is best shown by the Niagara Escarpment where the resistant dolomites and limestones stand high above the lowland to the north and east that is underlain mainly by shales.

Deep preglacial valleys in the bedrock are present at various places along the

Niagara Escarpment. One of the longest of these is the Dundas Valley that extends west from the tip of Lake Ontario for about 10 miles, at which point it becomes filled with drift. Other examples of preglacial bedrock valleys that have subsequently been filled are in the Georgian Bay area; these are the Beaver Valley extending from Thornbury to Flesherton and the drowned valleys that form the harbours at Owen Sound and Wiarton.

All these valleys were probably tributaries to a major valley that extended southeast from Georgian Bay to Lake Ontario and that formed a part of the preglacial Laurentian River System. From Montreal to Quebec City the preglacial Laurentian River was probably not far from the course of the present St. Lawrence River and was undoubtedly joined by tributaries that drained the uplands of the Canadian Shield and Appalachian hydrogeological regions.

Preglacial bedrock valleys that were probably part of the Laurentian River System have been detected beneath the drift by detailed seismic surveys. Such tributary valleys have been found in the area southeast of Rigaud, Quebec (Tremblay and Hobson, 1962), and in the Beauharnois and Chambly areas south of Montreal (Freeze, 1964). Further detailed investigations will be required throughout the St. Lawrence Lowlands before the full extent of these buried valley systems is known.

Productive aquifers have been found in buried valleys in places where substantial quantities of granular materials have been preserved. As an example, wells yielding several hundred gallons per minute have been developed from a sand aquifer contained in the buried valley near Rigaud, Quebec (Tremblay and Hobson, 1962). It is unlikely that aquifers will be found everywhere throughout the length of a particular buried valley; however, in areas underlain by relatively impervious strata the only potential aquifers may be those associated with a buried valley.

The total thickness of sediments removed by preglacial erosion is not known, but the present distributions of Palaeozoic rocks in southern Ontario indicates that the greatest thicknesses have been removed in areas adjacent to the Canadian Shield, and a similar relationship probably exists in the St. Lawrence Valley.

The development of joints within the rock masses, hence fracture permeability, may be attributed, in part, to the relief of stresses within the rocks resulting from the removal by erosion of overlying strata. Joints may also be produced by stresses imposed upon the rocks during their consolidation from sediments to rock and by the tectonic forces that have uplifted the rocks from a marine environment to their present elevations.

Surficial Deposits

During the Pleistocene epoch, the continental glaciers that moved over the St. Lawrence Lowlands further eroded the Palaeozoic rocks and substantially modified or obliterated the topography that existed in preglacial time. The Pleistocene epoch, which spans a period of about 1,000,000 years, consists of four glacial stages (Nebraskan (oldest), Kansan, Illinoian, and Wisconsin (youngest)) separated from one another by non-glacial intervals (Aftonian, Yarmouth, Sangamon). In the Mississippi Valley and in other parts of the midwestern United States, the sedimentary record of the Pleistocene epoch contains deposits that were laid down during each of

the glacial and interglacial stages. In southern Ontario, however, the oldest Pleistocene deposit is glacial till that has been assigned to the Illinoian stage. This till is overlain by non-glacial Sangamon sediments which in turn are overlain by glacial and non-glacial sediments deposited during substages of the Wisconsin stage. Throughout the St. Lawrence Valley, all the known Pleistocene glacial and non-glacial sediments were deposited during substages of the Wisconsin stage.

In the St. Lawrence Lowlands, therefore, the surficial deposits consist of glacial and non-glacial sediments deposited during the last glacial stage of the Pleistocene epoch, except for relatively small areas of Recent sediments that are generally associated with modern streams and rivers.

The aquifer potential of the Pleistocene deposits is for the most part dependent upon the grain-size characteristics of the sediments. Deposits composed mainly of sand and gravel are far superior as aquifers to deposits of silt and clay because of the greater ability of the coarser materials to transmit water. Most of the coarse-grained Pleistocene sediments are those that have been sorted, to varying degrees, by melt water that flowed from a waning glacier. These deposits are classified as stratified drift and include outwash, kames, eskers, and glacial spillway deposits. Other coarse granular deposits were formed along river terraces, along the shores of glacial lakes and inland seas, and in deltas that formed at the mouths of rivers that emptied into the lakes or inland seas. Till, which was deposited directly by the glacier with little or no sorting by running water, is generally a poor aquifer but may, in places, contain lenses and pockets of coarse sediments that form small, local aquifers.

Ontario

The glaciers that covered southern Ontario during the Wisconsin stage advanced from the centre of ice accumulation in Labrador. The ice probably moved as a broad stream up the St. Lawrence Valley into the lowlands of Lakes Ontario and Erie and as a lobe advancing south in the basin of Lake Huron. During the advance of the glaciers, much of the previously deposited glacial sediments were probably eroded and incorporated in the deposits of younger till. In some places, however, the advancing glaciers may have buried pre-existing, coarse, granular deposits, thereby creating potential aquifers beneath a covering of till. Most of the coarse, granular surficial deposits that are of significance as aquifers are related mainly to the events that took place during the northward recession of the glaciers, as it was during these phases that meltwater was most active in transporting and sorting sediments.

In the area north of Lake Ontario (Fig. 45, *in pocket*), the confluence of the two main glacial lobes is marked by an elongate kame moraine composed mainly of sand and gravel. The position of the confluence farther west is shown by similar kame deposits in the area between Orangeville and London.

The vast quantities of meltwater produced during the recessional phases of the glaciers were conveyed southward through spillways to lakes that formed in front of the ice masses. Coarse-grained sediments carried by the meltwater accumulated in the spillway channels as gravel terraces or as sheets of gravel and sand containing

some silt and clay. These spillway deposits are commonly thick and extensive and are important potential aquifers as well as sources of construction materials (Fig. 45).

The succession of lakes that occupied the lowland areas to the south of the receding ice masses had a complex history related to fluctuations of the ice front. Drainage outlets for the lakes were controlled by the position of the ice fronts—consequently the lake levels, and the extent and duration of the lakes were related to changes in ice front positions.

Sediments deposited in the offshore parts of the lakes were generally fine grained and therefore these deposits have little potential as aquifers. Shoreline or near-shoreline deposits of the lakes, however, such as beaches, beach bars, and spits, are commonly composed of sand or sand and gravel, and in places where these features are extensively developed they may be excellent aquifers. Most of the sand plains in southern Ontario, as shown on Figure 45, are glacial lake shoreline deposits. Delta deposits of sand and gravel that formed at the junction of the spillways with the glacial lakes may be important local aquifers but these deposits have not been shown separately on Figure 45. In southern Ontario east of the Frontenac axis, most of the sand deposits as shown on Figure 45 originated either as Champlain Sea beaches or as high level terraces formed during the early stages in the development of the Ottawa and St. Lawrence Rivers.

In order to be productive aquifers the surficial deposits must be sufficiently thick to extend into the saturated zone. In some areas of southern Ontario, in particular the area east of Lake Huron and south of Georgian Bay and the areas adjacent to the Frontenac axis, coarse granular surficial deposits do occur but they are probably of insufficient thickness to be productive aquifers. Average well yields from surficial aquifers range from 1 to 30 gpm (Fig. 45). The most productive surficial aquifers in southern Ontario are in the complex of spillway, beach, and kame deposits that lie in the area south from Elmira to Lake Erie.

Quebec

The unconsolidated Pleistocene sediments that mantle the bedrock of the St. Lawrence Valley are the result of (1) glaciation during the Wisconsin stage, (2) marine invasion during the recessional phases of glaciation, and (3) alluvial deposition during withdrawal of the seas from the area.

Till, deposited over the bedrock surface during the early phases of Wisconsin glaciation, is generally the oldest glacial deposit found in the area. In some places, however, as for example in the Bécancour area, stratified glacial lake sediments are found below the oldest till providing evidence for an earlier glacial episode. The distribution and extent of these earlier deposits is not known. The oldest till is overlain by stratified sediments of non-glacial origin indicating a recession of the early Wisconsin glacier. With a return to glacial conditions these older sediments were buried beneath a layer of till; then as the recession of this glacier began, deposits of stratified drift, kames and outwash, were accumulated by meltwater action.

In late Wisconsin time, recession of the glacier from the western side of the Appalachian highland permitted marine waters to invade from the east into the

St. Lawrence, Ottawa, and Champlain Valleys. The body of marine and brackish water formed by the invasion has been termed the Champlain Sea. Marine silts, clays, and minor sands that were deposited in the Champlain Sea form the thickest and most extensive sequence of surficial deposits throughout the St. Lawrence Valley.

Differential uplift of the St. Lawrence Valley, effected by continued northward recession of the glacier, caused the Champlain Sea to be drained from the lowland area. The sea was replaced in progressive stages first by estuarine, then by fresh-water lacustrine and fluvial ancestors of the present fresh-water drainage system. Erosion of pre-existing sediments and the deposition of extensive terrace deposits of sand attended the transition from the Champlain Sea to the present fluvial system.

The terrace and alluvial sands constitute some of the best aquifers in the St. Lawrence Valley. Wells that yield from 15,000 to 1,500,000 gpd have been developed in these deposits, and these wells serve as public water supplies for many of the small towns and villages in the area. Springs that issue from the contact of the terrace sands with the underlying marine clay are also used as sources of water supply for some communities. The marine silt and clay of the Champlain Sea sediments yield only small quantities of water to local users. In some places, however, marine beach deposits of sand and gravel are sufficiently thick and extensive to be used as aquifers that supply small communities.

Deposits of glacial outwash composed of sand and gravel are of more local occurrence than the terrace deposits, but where present they may yield large quantities of water. In some places, outwash deposits may be overlain by Champlain Sea sediments, and a careful examination of the local geology, accompanied by drilling, would be required to determine the presence of such deposits.

The till generally has a low potential as an aquifer but, as in other parts of the Lowlands, it may contain small pockets of sand and gravel that can be developed for local sources of water supply.

Anticosti Island

The trends of glacial striae and other streamline features indicate that the direction of ice movement during the last phase of Pleistocene glaciation was toward the southwest.

The bedrock surface above an elevation of 250 feet is covered by a thin veneer (from 2 to 6 feet) of till composed mainly of pebbles and cobbles of limestone, derived from the underlying bedrock, and contained in a matrix of silt or clay. Below elevations of 250 feet, the island has been subjected to marine inundation that has resulted in the development of a succession of wave-cut terraces mantled by stratified marine sediments. Deposits of sand and gravel occur at various places along the coast of the island particularly at the western end and near the mouths of rivers.

Although groundwater supplies have not been developed on Anticosti Island, it is probable that quantities of groundwater could be obtained from the sand and gravel deposits that locally attain an appreciable thickness if the need for this source of water should arise.

Groundwater Chemistry

The amount and type of chemical constituents contained in groundwater are, for the most part, functions of the chemistry of the materials encountered by the water in its passage through the groundwater flow system and the time required for the water to move from a recharge area to an area of discharge. Increases in the quantity of dissolved minerals contained in groundwater are generally encountered, therefore, with increases in well depth or in areas of natural groundwater discharge.

Throughout the St. Lawrence Lowlands, the shallow groundwater flow systems, from which most wells obtain water, closely coincide with the surface drainage systems (Fig. 46, *in pocket*). Thus the surface watersheds are areas of groundwater recharge and the streams and lakes receive groundwater discharge. Although a progressive increase in the total amount of dissolved minerals in groundwater could be expected from a comparison of the chemical analyses of groundwater obtained from wells ranging in location from an area of recharge to an area of discharge, the chemical data as shown on Figure 46 are not sufficient to demonstrate this relationship. In some areas, wells near the surface drainage divides produce waters that are more highly mineralized than those from wells in areas of probable groundwater discharge. This apparent anomaly probably results from the fact that the chemical analyses are of waters from a diversity of depths and thus represent portions of various flow systems.

Throughout most of southern Ontario, the main chemical constituents of groundwater are calcium and magnesium bicarbonates which reflect the chemistry of limestone and dolomite aquifers. A zone extending from Caledonia northwest to Southampton, however, is characterized by groundwater high in total solids of which the main constituent is calcium sulphate. It may be seen from Figure 44 that this area is underlain by the gypsiferous Salina Formation and thus the chemistry of groundwater in this area is controlled almost entirely by the local geology. Water containing sodium bicarbonate and hydrogen sulphide is commonly encountered in places where aquifers in black shales are utilized. Sodium chloride is also a common chemical constituent of bedrock waters and is often the dominant component in waters from deep bedrock aquifers.

Chemical analyses of groundwater derived from bedrock aquifers in the western part of the St. Lawrence Valley in Quebec also show the limestone aquifers to be high in bicarbonates of calcium and magnesium. Water from shale aquifers south of Montreal is high in sodium bicarbonate and many wells in the shale aquifer produce water with the unpleasant odour of hydrogen sulphide.

In some places throughout the St. Lawrence Valley, groundwater with a high content of sodium chloride is encountered. Some of the sodium chloride may be derived by the leaching of salt from the extensive deposits of marine clays in the area. Although groundwater containing substantial quantities of sodium chloride is generally undesirable for domestic water supplies, mineral spring waters of the sodium chloride type containing various trace elements are marketed for medicinal purposes. In 1962, Quebec supplied 98.8 per cent of the mineral waters in Canada, which had a total value of $211,825.

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Chapter V

CANADIAN SHIELD HYDROGEOLOGICAL REGION

by J. E. Charron

Introduction

The Canadian Shield hydrogeological region consists of the southern part of the Canadian Shield in Ontario and Quebec, which has a humid continental climate as contrasted with the subarctic climate of the vast area of Shield to the north (Fig. 7). The dividing line between the Canadian Shield hydrogeological region and the northern region approximates the southern limit of discontinuous permafrost west of James Bay and the change in forest cover east of James Bay.

This chapter describes the groundwater conditions in this region so far as they are known. No complete well inventory or groundwater studies have been published in the region, so the data available are scanty and the description is general, with no specific areas or problems discussed.

Physical Features

Nearly all the Shield under study is less than 2,000 feet above sea-level and the relief is commonly less than 200 feet, though along the southern edge it may be a few thousand feet. Rock exposures probably total less than 10 per cent of the surface area, and rock exposed has been well rounded by intense glaciation; this has left thousands of lakes connected by rivers with a completely disorganized drainage. These physical features of the Shield are considered as key factors in the control of groundwater distribution.

Climate

The climate of the Canadian Shield hydrogeological region is considered humid-continental. It is a climate with rain and snow. The mean temperature of the warmest month is above 50°F, whereas that of the coldest month is below 26°F. The wide range of temperatures that may be encountered in the course of a year can be illustrated by the extremes. The lowest temperature recorded is —60°F, the highest temperature is 105°F: a difference of 165°F.

The mean annual total precipitation varies from east to west. It is 44 inches in the east near the Gulf of St. Lawrence, but only 22 inches in the west near the beginning of the western plains. Much of this mean annual total precipitation falls as snow,

and again the variation holds true from east to west. There is a mean annual snowfall of 200 inches in the eastern part, which amounts to nearly half the total mean annual precipitation, whereas only 70 inches falls in the western part, or about one third of the total mean annual precipitation—assuming 10 inches of snow is equivalent to 1 inch of rain.

Throughout the Canadian Shield hydrogeological region there is ample precipitation to fulfill, at least theoretically, the recharge needs of all the groundwater basins. Normally, in the groundwater study the precipitation data are considered as one of the most important factors; but in this region, because it is ample, it is considered as one of the less important of the controlling factors in the hydrological cycle.

Geology

The geology of the Canadian Shield region is extremely complex and is explained in more detail in Economic Geology Report No. 1, *Geology and Economic Minerals of Canada.* Therefore, only those aspects that help to understand the groundwater situation are described briefly here.

Surficial Geology

Even though more than 90 per cent of the region is mantled with unconsolidated surficial deposits, these have not been mapped as fully as the bedrock, because the economic development of the region has been largely based on mining rather than agriculture, and because the dense forest cover makes rapid mapping difficult. However, a knowledge of these deposits is of utmost importance for an understanding of groundwater, as they include the most permeable materials and consequently have the greatest potential for large groundwater supplies.

Bedrock Geology

The Precambrian Shield region has been a stable mass since Precambrian time. It has not been folded by late mountain-building movements. Furthermore, erosion has worn it down to a point where it is one of the most solid masses of rock on earth.

The Archaean (early Precambrian) strata occur as altered volcanic and sedimentary rocks. These ancient formations were intruded by granitic rocks and occur today as bodies more than 100 miles long and many miles wide, which seem to be floating in a vast sea of granitic rocks.

Extensive bodies of comparatively undeformed volcanic and sedimentary rocks of Proterozoic age also exist. The oldest strata of this group have also been intruded by granitic rocks, while the youngest strata were intruded by diabase sills and dykes.

As shown on geological maps of Canada, most of the Shield area consists predominantly of granitic rocks, but as detailed mapping progresses, more of the ancient sedimentary and volcanic rocks are being discovered. At present, these rocks are mapped in detail only in the large mining areas of northeastern Ontario and northwestern Quebec. From the point of view of groundwater, the bedrock geology of the Shield is a very important factor, as is explained later.

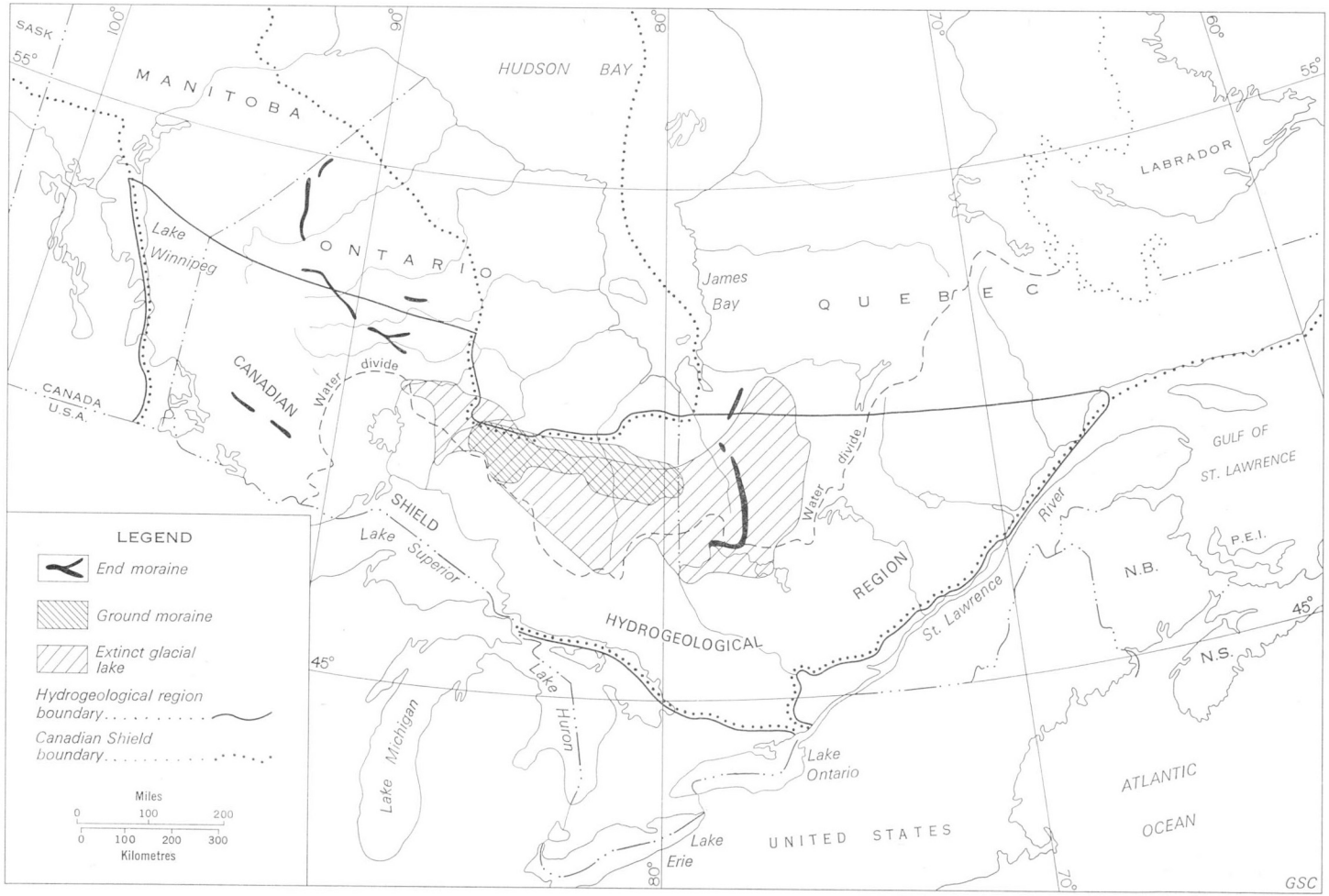


FIGURE 47. Areas of major surficial deposits in Canadian Shield hydrogeological region.

Hydrogeology

The Canadian Shield hydrogeological region includes parts of two large watersheds. The water divide line (Fig. 47) separates the Shield almost in half. The northern half forms part of the Hudson Bay watershed; the southern half forms part of the Atlantic Ocean watershed, and specifically the Great Lakes and St. Lawrence River watershed part of it. The water divide is almost parallel to the Great Lakes and St. Lawrence River system. Economically, the southern half is by far the most important, as most of the scattered Shield population lives in it.

There are 112 cities and towns with over 1,000 people (Table XI) in this part of the Shield, for a total population of some 900,000 (1961). Of these cities and towns, 46 have a lake as a source of water supply, 39 use a river, and 27 use groundwater. Assuming that most of the inhabitants living outside these communities make use of groundwater, then only some 20 per cent of the total population uses groundwater as a source of water supply. There are numerous reasons why groundwater is not more widely used in the Shield, but the principal one is the abundance of lakes and rivers. Because of the widely scattered population, most of these lakes and rivers have not yet been polluted and are not likely to be for some time. All told, a consumption of 97 million gpd is obtained from lakes, 42 million gpd from rivers, and 11 million gpd from groundwater (springs and wells). The largest demand registered is that of the pulp and paper industry. For example, the paper mill at Espanola, Ontario, uses 25 million gpd from the Spanish River. Industrial figures such as the example just given are not included in the daily consumption for the cities and towns mentioned.

Table XI

Water Supply of Communities over 1,000 Population in Canadian Shield Hydrogeological Region

Locality	Population	Consumption (gpd)	Remarks
FROM GROUNDWATER			
Almonte, Ont.	3,300	400,000	5 deep wells
Amos, Que.	5,450	500,000	springs
Blind River, Ont.	4,000	270,000	wells
Bourlamaque, Que.	3,000	250,000	springs
Bracebridge, Ont.	2,800	—	16 wells
Cadillac, Que.	1,000	150,000	2 wells
Capreol, Ont.	2,550	375,000	wells
Cochrane, Ont.	4,400	400,000	2 wells
Duparquet, Que.	1,500	85,000	
Englehart, Ont.	1,600	400,000	
Falconbridge, Ont.	1,400	3,000,000	deep wells
Gravenhurst, Ont.	3,100	450,000	3 deep wells
Larder Lake, Ont.	2,000	690,000	wells
Madoc, Ont.	1,500	—	deep wells
Manitouwadge, Ont.	2,400	117,000	2 deep wells
Marathon, Ont.	2,550	425,000	2 deep wells
Mattawa, Ont.	3,200	32,000	2 deep wells

Table XI *(cont'd)*

Water Supply of Communities over 1,000 Population in Canadian Shield Hydrogeological Region

Locality	Population	Consumption (gpd)	Remarks
FROM GROUNDWATER *(cont'd)*			
New Liskeard, Ont.	4,500	375,000	4 deep wells
Normandin, Que.	1,600	—	springs
St. Gabriel de Brandon, Que.	3,300	540,000	artesian wells
St. Raymond, Que.	3,500	125,000	springs
South Porcupine, Ont.	4,150	460,000	
South River, Ont.	1,050	38 000	springs
Tweed, Ont.	1,650	—	
Val d'Or, Que.	9,950	900,000	springs
Ville Marie, Que.	1,600	160,000	
FROM RIVER			
Coniston, Ont.	2,600	2,500,000	Wahnapitae R.
Deep River, Ont.	4,800	1,600,000	Ottawa R.
Dolbeau, Que.	5,600	975,000	Mistassini R.
Dryden, Ont.	5,500	520,000	Wabigoon R.
Espanola, Ont.	4,950	450,000	Mill—25,000,000
Gananoque, Ont.	5,100	448,000	St. Lawrence R.
Hearst, Ont.	2,400	350,000	Mattawishkwia R.
Jonquière, Que.	30,000	3,200,000	Aux Sables R.
Kapuskasing, Ont.	6,400	526,000	Kapuskasing R.
Ansonville and Iroquois Falls, Ont.	8,400	650,000	
Arnprior, Ont.	5,400	768,000	Madawaska R.
Arvida, Que.	14,200	3,382,000	Chicoutimi R.
Atikokan, Ont.	6,950	650,000	Atikokan R.
Brockville, Ont.	16,000	2,700,000	St. Lawrence R.
Buckingham, Que.	7,100	—	Lievre R.
Campbells Bay, Que.	1,000	25,000	Ottawa R.
Chicoutimi, Que.	52,150	3,200,000	Chicoutimi R.
Almaville, Que.	12,000	2,200,000	Grande Decharge R.
Maniwaki, Que.	6,350	150,000	Gatineau R.
Mountjoy township, Ont.	1,900	100,000	Mattagami R.
Pembroke, Ont.	15,600	1,500,000	Ottawa R.
Perth, Ont.	5,500	700,000	Tay R.
Pine Falls, Man.	1,200	235,000	Winnipeg R.
Port Alfred, Que.	8,500	1,300,000	Mars R.
Powassan, Ont.	1,050	50,000	Genesee Creek
Rainy River, Ont.	1,300	290,000	Rainy R.
Renfrew, Ont.	8,400	1,190,000	Bonnechere R.
Rivière du Moulin, Que.	4,600	350,000	Chicoutimi R.
Roberval, Que.	7,100	400,000	
Ste Anne des Monts, Que.	4,700	250,000	
St. Félicien, Que.	4,600	500,000	Ours R.
St. Jerome du Lac St. Jean, Que.	1,900	90,000	Lac au Sable R.
Sault Ste. Marie, Ont.	51,400	7,000,000	St. Mary's R.
Schumacher, Ont.	4,500	280,000	
Smooth Rock Falls, Ont.	1,150	1,000,000	Mattagami R.
Sturgeon Falls, Ont.	6,200	510,000	Sturgeon R.
Timmins, Ont.	28,000	2,300,000	Mattagami R.

Table XI *(conc'd)*

Water Supply of Communities over 1,000 Population in Canadian Shield Hydrogeological Region

Locality	Population	Consumption (gpd)	Remarks
		FROM LAKE	
Baie Comeau, Que.	6,650	1,000,000	Mill—13,000,000
Bancroft, Ont.	2,600	400,000	Clarke L.
Chambord, Que.	1,100	—	Belly L.
Chapleau, Ont.	3,700	300,000	Kebsquasheshing L.
Chibougamau, Que.	5,400	600,000	Gilman L.
Cobalt, Ont.	2,100	800,000	Sasaginaga L.
Copper Cliff, Ont.	3,800	1,500,000	Macdonald and Meatbird L.
Elliot Lake, Ont.	24,800	1,000,000	1 well
Flin Flon, Man.	12,000	550,000	Cliff and Trout L.
Fort Frances, Ont.	9,100	900,000	Rainy L.
Fort William, Ont.	43,500	8,230,000	Loch Lomond
Geraldton, Ont.	3,400	400,000	Reesor L.
Grand'Mère, Que.	15,250	2,100,000	Piles and Giguere L.
Haileybury, Ont.	2,550	310,000	Temiskaming L.
Hebertville Station, Que.	1,400	—	
Huntsville, Ont.	3,200	500,000	Vernon L.
Keewatin, Ont.	2,100	9,500	Lake of the Woods
Kenogami, Que.	11,300	1,300,000	Long L.
Kenora, Ont.	10,650	1,412,000	Lake of the Woods
Kirkland Lake, Ont.	20,000	4,000,000	McTavish and Victoria L.
La Sarre, Que.	3,700	125,000	1 deep well
La Tuque, Que.	12,200	50,000,000	Mill included
Malartic, Que.	7,200	470,000	1 well
Michipicoten, Ont.	3,500	1,000,000	Wawa L.
Mont Laurier, Que.	6,500	825,000	Thibeault and La Dam L.
Mont Rolland, Que.	1,700	100,000	Violon L.
Nipigon, Ont.	2,700	200,000	Helen L.
Nobel, Ont.	1,000	25,000	Georgian Bay
Noranda, Que.	11,400	1,000,000	Dufault L.
North Bay, Ont.	22,700	2,600,000	Trout L.
Parry Sound, Ont.	6,100	600,000	Georgian Bay
Port Arthur, Ont.	44,000	3,800,000	Superior L.
Rawdon, Que.	2,200	275,000	Vail L.
Red Lake, Ont.	2,200	—	Red L.
Red Rock, Ont.	1,800	365,000	Nipigon L.
Rouyn, Que.	18,000	800,000	
Ste. Agathe des Monts, Que.	5,500	1,200,000	Petit Lac des Sables
St. Jovite, Que.	1,650	—	Aqueduc and Briere L.
St. Sauveur des Monts, Que.	1,350	160,000	Milette L.
Schreiber, Ont.	2,100	200,000	Cook's L.
Shawinigan, Que.	43,600	4,200,000	La Pêche L.
Sioux Lookout, Ont.	2,600	600,000	Pelican L.
Sudbury, Ont.	76,800	4,400,000	2 deep wells
Temiskaming, Que.	2,800	—	Temiskaming L.
Thessalon, Ont.	1,700	450,000	Huron L.

Information from: Canadian Municipal Utilities, Annual Directory, February, 1961

Surficial Deposits

So little information on this subject is available in the Canadian Shield hydrogeological region that very little can be said. Sand and gravel aquifers, however, should be the best sources in the Shield, if large quantities of groundwater are required.

As more Pleistocene mapping is done, more areas favourable as a source of groundwater will be delineated. At present, areas such as the large clay belt that was once covered by glacial lake Barlow-Ojibway, and areas covered by ground moraine consisting basically of clayey till, can be considered as poor groundwater areas. Some good aquifers occur in granular materials beneath the glacial lake clays, but intensive exploration will be required to determine the extent of such aquifers and the best means of finding and developing them.

Bedrock

Most of the available information on the occurrence of groundwater in bedrock is derived from the experience of the mining industry. As nearly all the rocks in this region are crystalline, almost all the groundwater occurs in fractures; and as fractures are found everywhere, but vary widely in distribution and size, so groundwater is found everywhere, but varies widely in quantity. Figures presented in the accompanying tables show that, although small supplies can be generally obtained, large supplies suitable for industry or community water supplies will be found only in the exceptional areas where structures are favourable, or can be obtained at great expense by developing large well-fields.

Crystalline rocks commonly have a very low porosity, in the range of 1 to 0.5 per cent. On the other hand, unaltered sedimentary and volcanic rocks have a much higher porosity and permeability. Particularly, volcanic rocks such as basalts, tuff (average porosity 30%), and pumice (maximum porosity 87%) have high porosity and permeability, which facilitate the circulation of groundwater.

Geological studies of the Shield area show that it has been deeply eroded, and consequently nearly all rocks now near surface have been subjected to deep burial and have undergone repeated orogeny. Due to this history, any original porosity and permeability have been lost, and only fracture porosity and permeability are left. In addition, the glaciers of the ice age have removed all but a few small remnants of any porous weathered zone that may have existed, and there has not been sufficient time since to develop any appreciable weathered zone. We are dealing with one of the most solid masses of rock on earth.

Mining experience shows that, although fractures are present and numerous at depth, they also are closed more tightly than those near surface and consequently carry less water. Unfortunately, little quantitative data are available on the decrease in amount of groundwater with depth, and only general conclusions can be drawn.

Quantitative Data

Data collected in 1962 by the Mineral Economics Section, Mineral Resources Division, Department of Mines and Technical Surveys, on the water pumped by mines in this region are presented in Table XII. In using these figures, it must be remembered

that the actual groundwater pumped is less than that shown, as the total water pumped per day includes any water piped undergound from surface to be used for operations such as drilling or washing, but excludes water used in back-filling.

A large mine may have many hundreds of miles of drifts, crosscuts, and raises, and thousands of diamond-drill holes made in all directions from these openings and from the large mined-out areas. These numerous openings may extend from a depth of 100 feet to several thousands of feet and very efficiently tap the groundwater over the full area of the mine. They also draw in water, as a well does, from a cone of depression of unknown size.

Of the 53 mines listed in Table XII, Hollinger Consolidated Gold Mines Limited pumps the largest amount of water—4,100,000 gpd in the springtime and a normal average of 1,900,000 gpd. The average of the 53 mines is 275,000 gpd, or some 190 gpm. If these efficient collectors can produce only this amount of water, it is apparent that an individual well of 6 inches in diameter and a few hundred feet in depth will tap far fewer fractures and can be expected to produce much less water, unless it happens to intersect a particularly open and extensive fracture system. It would require a large well-field to produce water equivalent to a mine.

Fault zones are usually the zones yielding the largest quantities of groundwater, but this depends on whether the fractures in the fault zones are open or whether they are sealed, and a fault can also be a barrier to groundwater movement.

A good example of how impermeable the intrusive rocks of the Precambrian Shield can be is the value given for Sullivan Consolidated Mines Limited. The underground workings are situated beneath Lake Dubuisson, extending to within some 100 feet of the lake bottom, and yet the amount of water pumped daily at that mine amounts to only 200,000 gpd.

The uranium mines at Elliot Lake show that less water is encountered with increased depth, because the fractures or fissures are small and tight. The three mines used here as examples are Nordic, Denison, and Milliken. Nordic is the shallowest mine, at 1,400 feet; the depth of Denison and Milliken is 2,465 and 3,000 feet, respectively. Denison is the largest producer, at 6,000 tons daily. Nevertheless, Nordic pumps 400,000 gpd out of the mine, as compared to very little water pumped out by the other two mines, which are deeper. As the smallest amount recorded in Table XII is 20,000 gpd, it is assumed that the amount of water pumped out of Denison and Milliken is less than that figure. These mines are mining a flat-lying orebody, and all their openings except the shafts occur at 1,000 to 2,500 feet below surface or lake bottom, so they give a sample of conditions at depth, while most other mines give a sample extending from near surface to depth.

Geologically speaking, most orebodies in the Shield occur within the sedimentary or volcanic rocks, or at their contact with the granitic rocks, but not within the granitic rocks themselves. Yet it was said earlier in this chapter that these same sedimentary and volcanic rocks are the most porous of the crystalline rocks. Therefore, the small amount of water pumped out of these 53 mines in the Shield confirms the fact that these supposedly porous rocks are not so porous simply because of their deep-seated character.

Table XII

Water Pumped from Mines in the Canadian Shield Hydrogeological Region

Mine	Location	Ore Mined	Production Tons of OreDaily	Major Mining Method	Depth Feet	Water Pumped gpd	Remarks
Geco Mines Ltd.	Manitouwadge, Ont.	Cu–Zn	3,500	open stoping	2,460	280,000	
Willroy Mines Ltd.	Manitouwadge, Ont.	Cu–Zn	1,000	open stoping	2,200	100,000	
Campbell Red Lake Mines Ltd.	Balmerton, Ont.	Au	700	shrinkage	3,250	210,000	
Cochenour Willans Gold Mines Ltd.	Balmerton, Ont.	Au	300	shrinkage	2,210	168,000	
Dickenson Mines Ltd.	Balmerton, Ont.	Au	500	shrinkage	3,365	145,000	
Madsen Red Lake Gold Mines Ltd.	Madsen, Ont.	Au	850	cut and fill	4,075	109,000	
McKenzie Red Lake Gold Mines Ltd.	McKenzie Island, Ont.	Au	250	open stoping	1,950	100,000	
East Sullivan Mines Ltd.	Val d'Or, Que.	Cu	3,000	open	4,000	250,000	
Manitou-Barvue Mines Ltd.	Val d'Or, Que.	Zn–Ag	1,300	cut and fill	4,200	120,000	
Bevcon Mines Ltd.	Perron, Que.	Au–Ag	650	open	2,250	90,000	
Lamaque Gold Mines Ltd.	Bourlamaque, Que.	Au–Ag	2,000	cut and fill	3,600	200,000	
Sullivan Consolidated Mines Ltd.	Sullivan, Que.	Au–Ag	650	shrinkage	3,100	200,000	under lake
Sigma Mines (Quebec) Ltd.	Bourlamaque, Que.	Au–Ag	1,200	shrinkage	4,050	little	
East Malartic Mines Ltd.	Norrie, Que.	Au	1,500	open	4,950	500,000	
Barnat Mines Ltd.	Malartic, Que.	Au	1,600	open	2,400	350,000	
Canadian Malartic Gold Mines Ltd.	Malartic, Que.	Au–Ag	1,000	cut and fill	1,625	150,000	
Henderson (Campbell Chibougamau Mines Ltd.)	Chibougamau, Que.	Cu	1,200	cut and fill	700	3,000,000	lake inflow—fissure
Copper Rand Chibougamau Mines Ltd.	Chibougamau, Que.	Cu	1,000	cut and fill	1,910	510,000	
Portage Island (Copper Rand Chibougamau Mines Ltd.)	Chibougamau, Que.	Cu	500	cut and fill	850	440,000	
Jaculet (Copper Rand Chibougamau Mines Ltd.)	Chibougamau, Que.	Cu	200	shrinkage	600	200,000	
Merrill Island Mining Corporation Ltd.	Chibougamau, Que.	Cu	?	cut and fill	2,275	95,000	
Main Mine (Campbell Chibougamau Mines Ltd.)	Chibougamau, Que.	Cu	?	cut and fill	3,650	little	
Bicroft (Macassa Gold Mines)	Bancroft, Ont.	U	1,300	shrinkage	1,650	350,000	
Faraday Uranium Mines Ltd.	Bancroft, Ont.	U	1,500	shrinkage	1,455	310,000	
Nordic (Rio Algom Mines Ltd.)	Elliot Lake, Ont.	U	4,500	room and pillar	1,400	400,000	
Stanrock Uranium Mines Ltd.	Elliot Lake, Ont.	U	4,000	panel stoping	2,300	20,000	

Milliken (Rio Algom Mines Ltd.)	Elliot Lake, Ont.	U	4,000	pillar room and pillar	3,000	little	
Wright-Hargreaves Mines Ltd.	Kirkland Lake, Ont.	Au	420	shrinkage	8,100	200,000	
Lake Shore Mines Ltd.	Kirkland Lake, Ont.	Au	450	square set	8,075	200,000	
Tech-Hughes Gold Mines Ltd.	Kirkland Lake, Ont.	Au	350	shrinkage	6,100	150,000	
Macassa Gold Mines Ltd.	Kirkland Lake, Ont.	Au	420	shrinkage	6,300	40,000	
Kerr-Addison Gold Mines Ltd.	Virginiatown, Ont.	Au	4,000	cut and fill	6,000	288,000	
Upper Canada Mines Ltd.	Dobie, Ont.	Au	550	cut and fill	4,900	75,000	
Stobie (International Nickel Co. of Canada)	Sudbury, Ont.	Ni–Cu	—	blasthole	3,000	500,000	
Frood (International Nickel Co. of Canada)	Sudbury, Ont.	Ni–Cu	—	square set	3,600	300,000	
Murray (International Nickel Co. of Canada)	Sudbury, Ont.	Ni–Cu	—	blasthole	3,000	300,000	
McKim (Falconbridge Nickel Mines Ltd.)	Sudbury, Ont.	Ni–Cu	300	blasthole	1,175	300,000	
Crean Hill (International Nickel Co. of Canada)	Sudbury, Ont.	Ni–Cu	—	blasthole	2,000	210,000	
Fecunis Lake (Falconbridge Nickel Mines Ltd.)	Onaping, Ont.	Cu–Ni	3,000	cut and fill	3,775	85,000	
Hardy Boundary (Falconbridge Nickel Mines Ltd.)	Onaping, Ont.	Cu–Ni	1,500	cut and fill	3,050	70,000	
Creighton (International Nickel Co. of Canada Ltd.)	Creighton, Ont.	Ni–Cu	—	caving	6,600	370,000	
East Falconbridge Nickel Mines Ltd.	Sudbury, Ont.	Cu–Ni	2,850	cut and fill	6,040	280,000	
Garson (International Nickel Co. of Canada Ltd.)	Garson, Ont.	Ni–Cu	—	cut and fill	5,000	210,000	
Hollinger Consolidated Gold Mines Ltd.	Timmins, Ont.	Au	3,000	cut and fill	5,450	1,900,000	max-4,100,000-spring
McIntyre Porcupine Mines Ltd.	Schumacher, Ont.	Au	3,000	cut and fill	8,000	800,000	
Aunor Gold Mines Ltd.	Timmins, Ont.	Au	760	square set	2,900	430,000	includes water from Buffalo Ankerite
Preston Mines Ltd.	South Porcupine, Ont.	Au	700	cut and fill	4,180	220,000	
Delnite Mines Ltd.	Timmins, Ont.	Au	500	cut and fill	5,250	75,000	
Paymaster Consolidated Mines Ltd.	South Porcupine, Ont.	Au	500	shrinkage	4,075	26,000	
Pamour Porcupine Mines Ltd.	Pamour, Ont.	Au	1,800	blasthole	3,100	576,000	
Broulan Reef Mines Ltd.	South Porcupine, Ont.	Au	460	cut and fill	2,560	300,000	
Hallnor Mines Ltd.	Pamour, Ont.	Au	375	cut and fill	3,700	120,000	

Information from: Mineral Economics Section, Mineral Resources Division

Quality of Water

Crystalline rocks are formed of minerals of low solubility; therefore, the water should have a low total dissolved solids content. In general, the groundwater in the Shield is similar to the acidic surface waters and has the same low pH. Due to the disorganized drainage and multitudinous bogs and marshes, the surface waters are mostly dark brown and high in organic matter. It is notable that, although both the organic content and stream stage fluctuate widely, the chemical content remains very constant, indicating that the groundwater discharged to streams is similar in composition to the surface water, and strongly suggesting that groundwater flow systems are comparatively short.

Strongly acidic waters are also encountered in some mining areas, especially where groundwater penetrates sulphide orebodies. In most gold mines the groundwater is potable, but such is not always so in the base metal mines.

Generally speaking, the groundwater found in the Shield should be of the bicarbonate type, simply because recharge can occur anywhere. The base exchange index should be negative; that is, the sodium (Na) ion should be greater than the chloride (Cl) ion. In granitic rocks the Mg/Ca ratio should be the same as the Mg/Ca ratio in the granitic rock itself.

Summary

The surficial waters of rivers and lakes satisfy most requirements in the Canadian Shield hydrogeological region at present.

From the little knowledge available, it would seem that the crystalline rocks can be considered as poor aquifers, whereas surficial deposits such as end moraines and outwash deposits would seem to be the best aquifers.

Generally speaking, hydrogeological studies of the Shield region are lagging behind those of other regions because requirements have not demanded them. Yet studies of this type will be required for future development.

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Chapter VI

INTERIOR PLAINS HYDROGEOLOGICAL REGION

by P. Meyboom

Hydrogeology

The Canadian Interior Plains lie between the Rocky Mountains and the Canadian Shield and extend from the American Great Plains to the Arctic Ocean. The Interior Plains hydrogeological region is the part of these plains south of the southern limit of discontinuous permafrost.

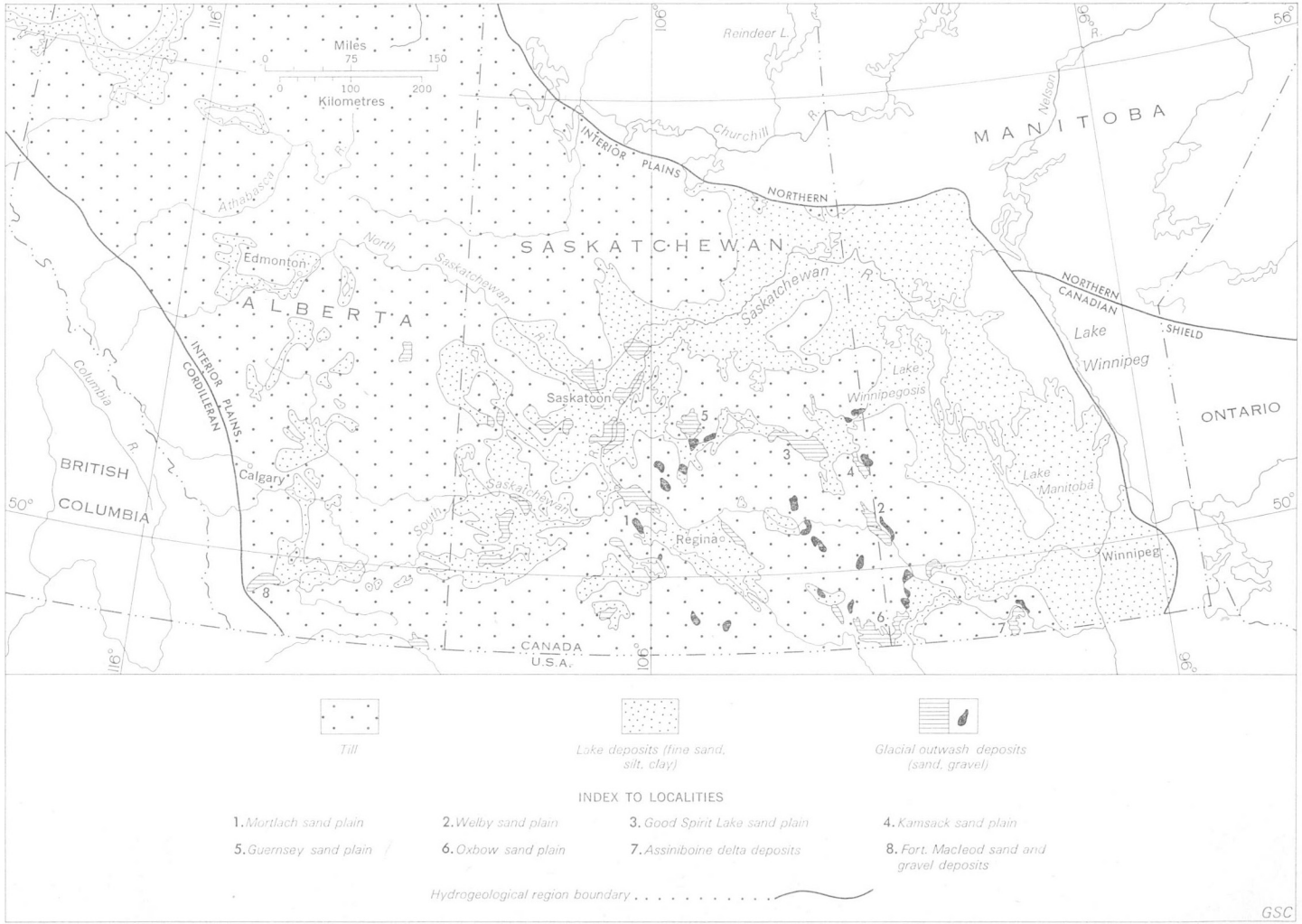
The entire plains area is underlain by nearly horizontal strata of Palaeozoic, Mesozoic, and Tertiary age, and is almost entirely covered by Pleistocene deposits. As a result of noticeable differences in hydrologic characteristics, much of the published information on the hydrogeology of the Interior Plains region has been based on a differentiation between glacial drift and bedrock; this division has been adhered to in the following description of the materials that constitute these two important hydrogeological entities.

Surficial Deposits

Nature and Distribution of Surficial Deposits

The surficial deposits of the plains fall into three categories: till, which is an unsorted mixture of sand, silt, clay, and boulders; lacustrine sediments, mainly silt and clay; and outwash deposits, whether deltaic sands and gravels or blanket outwash sands. The map of the regional distribution of these sediments (Fig. 48) shows that 60 per cent of the plains is covered by till, 40 per cent by lake deposits, and less than one per cent by outwash deposits.

The *till* is commonly calcareous, and its texture ranges from loam and clay loam to sandy clay loam (Fig. 49). Although the prairie tills as a whole have a remarkably uniform texture, Gravenor and Bayrock (1961) found that the tills of southeastern Alberta are less argillaceous than those of eastern Alberta, which in turn are less clayey than those of north-central and northwestern Alberta. The difference in clay content ranged from 30 per cent in southeast Alberta to 60 per cent in northwest Alberta.


Miles
0 75 150
0 100 200
Kilometres
Reindeer L.
Churchill R.
Nelson R.
MANITOBA
INTERIOR PLAINS
NORTHERN
SASKATCHEWAN
Athabasca
Edmonton
North Saskatchewan R.
ALBERTA
Saskatchewan R.
NORTHERN CANADIAN SHIELD
Lake Winnipeg
Lake Winnipegosis
ONTARIO
Columbia R.
INTERIOR PLAINS CORDILLERAN
Saskatoon
Calgary
BRITISH COLUMBIA
South Saskatchewan R.
Lake Manitoba
Regina
Winnipeg
50°
CANADA
U.S.A.
Till
Lake deposits (fine sand, silt, clay)
Glacial outwash deposits (sand, gravel)
INDEX TO LOCALITIES
1. Mortlach sand plain
2. Welby sand plain
3. Good Spirit Lake sand plain
4. Kamsack sand plain
5. Guernsey sand plain
6. Oxbow sand plain
7. Assiniboine delta deposits
8. Fort. Macleod sand and gravel deposits
Hydrogeological region boundary
GSC

The colour of the oxidized till is brown, whereas the unoxidized till is bluish grey. A characteristic clay-mineral distribution of prairie till shows 60 per cent montmorillonite, 20 per cent illite, and 20 per cent kaolinite. Most of the till is derived from local bedrock shales and sandstones.

The overall average thickness of till on the prairies is about 75 feet, but the average thickness in each province increases markedly from Alberta to Manitoba.

Near the surface, the permeability of till (Fig. 49) ranges from 0.3 to 0.5 inch per hour, but it decreases by a factor 10 or more in the interval from 10 to 100 feet below surface. For instance, "vertical permeabilities" obtained by Lissey (1962) from pumping tests on a leaky artesian aquifer underlying 140 feet of till and 44 feet of lacustrine clays in the Regina area varied from 0.002 inch per hour to 0.05 inch per hour (0.02 g/d/ft^2 to 0.6 g/d/ft^2). Tritium determinations on groundwater in till of the Missouri Coteau in central Saskatchewan indicated an average permeability of 0.014 inch per hour to a depth of 30 feet (Mawson, 1964).

The bulk of the *lacustrine deposits* is made up of clay (Fig. 49), which is characteristically calcareous, montmorillonitic, and unoxidized. The stratification of the lake deposits varies from perfect varves to crude bedding. The lacustrine deposits occur in extensive basins, once occupied by proglacial lakes such as Lake Edmonton, Lake Regina, and—the largest of all—Lake Agassiz in southern Manitoba. Gravenor and Bayrock (1961), as well as Elson (1961), pointed out that the distribution of lake deposits is a consequence of the regional northeast topographic slope and the north and northeastward direction of ice retreat. As the ice withdrew, ever lower outlets were uncovered, and consequently the distribution of lacustrine sediments shown in Figure 48 represents a composite picture of all successive lake stages that existed during the entire ice retreat.

The thickness of the lake deposits in the deepest parts of the former basins is about 40 feet (35 feet at Edmonton, 40 feet at Regina, 20 to 40 feet at Winnipeg), but diminishes toward the peripheries of the lake basins.

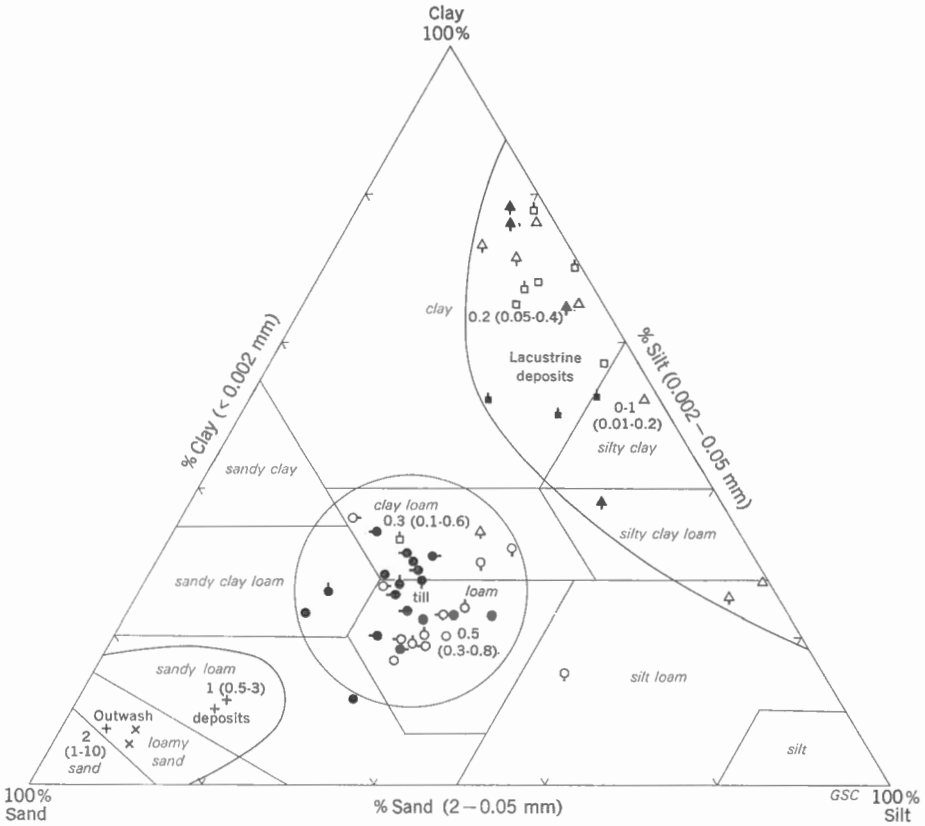
The permeability of the lake sediments is low (Fig. 49), ranging from 0.1 to 0.2 inch per hour near the surface to 0.001 to 0.002 inch per hour in the interval from 10 to 100 feet below surface.

Outwash deposits are found along the shores of the former glacial lakes and as alluvial fans at the mouth of meltwater channels. They consist chiefly of well-sorted sand and fine gravel, and crossbedding or channel-fill structures are common. The thickness of some of the more extensive sand deposits may be up to 50 feet (Welby sand plain) or 100 feet (Carberry sand plain), but is commonly less than 15 feet.

The permeability of these sediments is about 1 inch to 10 inches per hour (Fig. 49); these permeabilities do not seem to decrease as much with depth as they do in till and lake sediments.

Surficial Deposits as Aquifers

An estimated 60 per cent of all farm water supplies in the prairie provinces is obtained from the till, and from this standpoint the importance of the bulk of the surficial deposits should not be underestimated. However, most of these shallow wells yield less than 5 gpm and cannot be relied on for large sustained yields.



Conversion table for permeability units

	metre/day	inch/hr	g/d sq ft
1 metre/day	1	1.63	20.44
1 inch/hr	0.61	1	12.45
1 g/d sq ft	0.049	0.079	1

Average permeability (hydraulic conductivity) and ranges of permeability in inches/hour of principal Pleistocene deposits
Note: The values shown pertain to a depth of 0 – 10 feet. Below that depth the permeability decreases by a factor 10 in the interval from 10 – 100 feet below surface.

GRAINSIZE AND SOURCES OF INFORMATION

MANITOBA

- □ *LAKE AGASSIZ CLAY, Winnipeg area, from: Preliminary Soil Mechanics aspects of the Red River floodway, by J. Mishtak, 1962*
- ○ *NON-CEMENTED TILL, Winnipeg area (Mishtak, 1962)*
- ▲ *LAKE AGASSIZ CLAY, from: Reconnaissance Soil Survey of Winnipeg, Morris and Carberry Map sheets by W. A. Ehrlich et al. 1953 and 1957*
- \+ *OUTWASH SANDS, Carberry sand plain, (Ehrlich et al, 1957)*
- ● *TILL, south-western and south-central Manitoba, from: Reconnaissance soil survey of south-western and south-central Manitoba, by J. H. Ellis and Wm. H. Shafer, 1940 and 1943*
- × *SAND DEPOSITS of the Assiniboine delta (Ellis and Shafer, 1943)*
- ◊ *TILL, Virden area, central Manitoba, from: Reconnaissance Soil Survey of Rossburn and Virden map sheets by W. A. Erlich et al. 1956*

SASKATCHEWAN

- ◆ *TILL, Qu'Appelle area, from: Geology and groundwater resources of the Qu'Appelle area, Sask. by E. A. Christiansen, 1960*
- ○ *TILL, Regina area, from: Geology and groundwater resources of the Regina area by E. A. Christiansen, 1961*
- ▲ *LAKE REGINA CLAY. (Christiansen, 1961)*
- ● *TILL, (Upper Till) Wynyard area, from: Geology and groundwater resources of the Wynyard area by E. A. Christiansen, 1963, and J. E. Greer*
- ○ *TILL, Rosetown area, from: Surficial geology of the Rosetown map area, by J. S. Scott*
- △ *LACUSTRINE DEPOSITS, Rosetown area (Scott)*

ALBERTA

- ● *TILL, Edmonton area, from: Soil survey of Edmonton sheet, by W. E. Bowser et al, 1962*
- ▲ *LACUSTRINE DEPOSITS, Edmonton area (Bowser, 1962)*
- ○ *TILL, Grande Prairie area, from: Soil Survey of the Grande Prairie and Sturgeon Lake sheets, by Wm. Odynsky et al. 1956*
- △ *LACUSTRINE DEPOSITS, Grande Prairie area, (Odynsky, 1956)*
- ● *TILL, average for East-Central Alberta, from: Surficial Geology of the Edmonton district, by L. A. Bayrock and G. M. Hughes, 1962*

FIGURE 49. Grain-size distributions of principal Pleistocene deposits of the Interior Plains.

A noticeable and encouraging exception to this generalization are the intertill aquifers near the city of Regina, which have a combined safe yield of nearly 20 million gpd (Lissey, 1962). This example indicates that till deposits should not be a priori rejected as a possible source for water supplies that require over 1 million gallons per day. Thick and extensive intertill sandbeds may be present in other places, particularly in Saskatchewan and Manitoba, where the glacial drift is thicker than in Alberta and where there is ample evidence of more than one till sheet. Wickenden (1931) mapped intertill sands and gravels up to 225 feet thick in a tributary valley near the mouth of Swift Current Creek.

Most of the lacustrine sediments are aquicludes rather than aquifers, but Elson (1961) stated that thick sandy varves at the base of Lake Agassiz deposits are sufficiently permeable locally to be good aquifers.

Although the outwash deposits are limited in extent, their aquifer potential may be most important on the prairies, particularly in conjunction with artificial recharge, as was demonstrated in Alberta by Farvolden (1963b). On the other hand, on the Carberry sand plains in Manitoba, industrial wells that operate without artificial recharge produce over 1 million gallons per day without apparent depletion, owing to the circumstance that their use is restricted to a few months per year.

Influence of Landforms and Surficial Deposits on Groundwater Regime

The low permeability of the bulk of the surficial deposits affects the groundwater regime in various ways, depending on the land forms and on the part of the flow system that is being considered. Most of the till on the plains occurs as ground moraine, which has a flat to gently rolling surface with a relief of 5 to 30 feet. Owing to the low permeability of the till and the presence of more permeable strata in the underlying bedrock, groundwater in till moves downward over large areas of the plains. Annual recharge in these areas is about 0.5 inch (*see* section on water balance).

Hummocky moraine, which is composed of till knobs and kettles, as well as linear till ridges and small rimmed depressions, is a typical landform of many topographically high areas. Because of the accumulation of melting snow in the depressions, hummocky moraine would seem to be an optimum recharge environment. However, infiltration is slow, and many of the ponds contain water for a sufficient length of time during the growing season to permit the growth of willows and poplars around their edges. These phreatophytic trees and shrubs not only intercept much of the water that otherwise would have infiltrated, but also divert and even reverse the groundwater flow in their vicinity, as a result of the cone of depression that originates around their roots.

The effect of the lacustrine deposits in areas where groundwater discharge is restricted to deeply incised river valleys is to retard recharge over the lake plain. Wherever the discharge areas are of greater areal extent, the lacustrine clays may become 'classical' aquicludes, as can be seen in the artesian areas along the southern flank of the Moose Mountain upland, as well as in the Red River artesian basin.

Bedrock Valleys

Bedrock-Valley Systems in Alberta and Saskatchewan

Much of the drainage system that existed prior to or during the last glaciation of the prairies has been reconstructed from subsurface information. The bedrock-channel systems that were discovered in Alberta (Farvolden, 1960, 1963; Stalker, 1961) and Saskatchewan (Christiansen, 1962) are made up of channels that existed prior to glaciation (preglacial valleys) and those that were cut into bedrock during the Pleistocene (glacial or interglacial bedrock valleys). Both types of valleys may be either partly or completely filled with glacial drift. Those that are completely filled have been the subject of intensive searches on the plains, for many of the buried valleys contain important aquifers.

Both Farvolden and Christiansen advanced the argument that, owing to the very low gradients of the preglacial streams (of the order of 0.0009), significant amounts of sand and gravel can be expected only near those ancestral headwaters where Tertiary and pre-Pleistocene sandstones and conglomerates were exposed. Figure 51 shows those segments of preglacial streams in Saskatchewan that, according to Christiansen (1962), may contain extensive deposits of glacial and preglacial sands and gravels.

With respect to channels filled with coarse glacial material, Christiansen pointed out that in Saskatchewan the preglacial channels in which rivers flowed south, carrying material away from the glacier, are more likely to contain extensive deposits of sand and gravel than those in which the rivers flowed north, in the same direction as the ice retreat. Stalker (1961) stated that interglacial valleys in Alberta are characterized by till fill and a lack of gravel at their base.

With regard to their base level, there is a very significant difference between bedrock channels in Alberta and those in Saskatchewan. Farvolden (1963a) stated that "uplift and erosion that began in middle Tertiary time has not changed significantly, and the regional base level of the streams in Alberta is as low as, or lower than, it has ever been since the uplift of the Rocky Mountains." Since Oligocene time the geological record in Alberta has been chiefly that of degradation.

On the other hand, Christiansen (1962) concluded that "in Saskatchewan the base level of the preglacial valleys ranges from 100 to 300 feet below that of the present drainage, (which) suggests that there has been a continuous rise in base level since preglacial time." The rising base level in Saskatchewan has protected the previously deposited sand and gravel from subsequent erosion, whereas in many places of Alberta, where the bedrock channels coincide with the present-day valleys, erosion has removed most of the alluvium that may have been deposited in the channel at one time. Even in the case of buried valleys in Alberta, the base of the bedrock valley may be above the base of the present valley (for instance, the Oldman River at Taber and Lethbridge), and groundwater has been drained out of the buried stream deposits by the nearby stream. The rise of Alberta, relative to Saskatchewan, thus clearly favours the hydrogeological circumstances for buried stream-channel aquifers in Saskatchewan.

Hydrologically, the buried valley that contains coarse deposits may show as a depression in the water table or in the piezometric surface, indicating that the buried

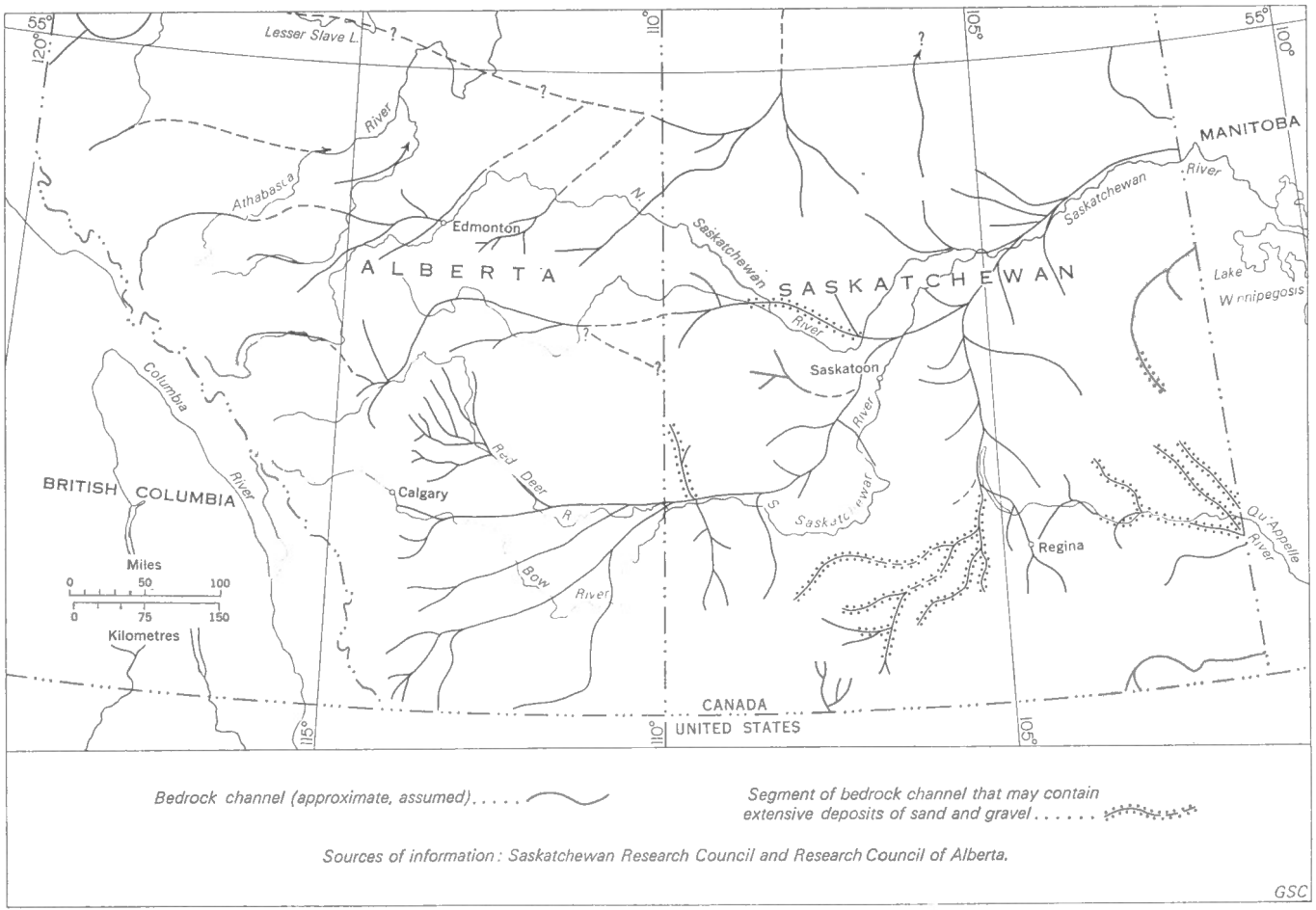


FIGURE 50. Bedrock channels of southern Alberta and Saskatchewan.

valley acts as a subdrain. LeBreton (1963c) described this drain effect from the Vegreville channel in eastern Alberta.

Hydrochemically, bedrock channels may be characterized by groundwater with less total dissolved solids than groundwater in the surrounding till. LeBreton and Jones (1962) drew attention to such a correlation along the Red Deer and Hanna channels in Alberta.

Aquifers in Buried Valleys

Buried bedrock valleys tend to be broad, commonly between 2 and 10 miles wide with gently sloping sides (Stalker, 1961), and 100 to 300 feet deep. At a few locations, notably at Medicine Hat (Alberta) and Regina (Saskatchewan), the coarse deposits at the base of such channels are exploited successfully as aquifers. A summary of the properties of these aquifers, as well as of buried-valley aquifers elsewhere on the prairies, is given in Table XIII. A map showing the bedrock valleys in Alberta and Saskatchewan is shown in Figure 50.

Table XIII

Hydrological Characteristics of Aquifers in Buried Bedrock Valleys of Alberta, Saskatchewan, and Manitoba

Depth below surface (in feet)	Saturated thickness of gravel (in feet)	Transmissibility g/d/ft	Coefficient of storage	Location
10	30	120,000	water table	Medicine Hat[1]
420	35	90,000	2.5×10^{-4}	Regina[2]
300	68	42,000	5×10^{-4}	Estevan[3]
200	up to 200	700,000	2×10^{-4}	Winkler[4]
120	15	5,700	1.2×10^{-4}	Hanna[5]

[1]Meyboom, P., 1963. *Res. Council, Alberta*, Bull. 12
[2]Lissey, A., 1962. City of Regina Eng. Dept., Hydrology Division, Rept. No. 1
[3]Christiansen, E. A., and Parizek, R.R., 1961. *Sask. Res. Council*, Geol. Div., Circular 1
[4]Charron, J. E., 1964. *Geol. Surv. Can.*, Paper 63–43
[5]Kunkle, G. R., 1962. *Res. Council, Alberta*, Prelim. Rept. 62–3

Bedrock Formations

Cretaceous and Tertiary Sediment Associations of the Western Canada Sedimentary Basin

Cretaceous and Tertiary sediment associations make up almost all the near-surface bedrock formations in the Western Canada Sedimentary Basin. They occur in a westward-dipping homocline, which extends 700 miles from the Manitoba escarpment to the foothills of the Rocky Mountains. The thickness of the Upper Cretaceous and Tertiary section ranges from 1,600 feet in Manitoba to more than 6,000 feet in the Alberta geosyncline (Fig. 51).

The geological history of the plains during and after the Cretaceous period has been analyzed by Webb (1954) and is summarized in the following paragraphs.

During early Cretaceous times the southwestern part of the Canadian plains witnessed wide transgression of continental deposits derived from the newly uplifted Cordilleran geanticline (Blairmore Formation) and from a landmass at the northeast in the present shield (Grand Rapids, McMurray, and Swan River Formations), whereas in the north and northwest, broad shallow seas transgressed the shelf from the north. Early Cretaceous passed into late Cretaceous with widespread gentle subsidence when the marine sedimentation of the Colorado stage commenced.

During Colorado time a thick section of marine shale accumulated on the plains area, presently constituting the Colorado Group in Alberta and its stratigraphic equivalents in eastern Saskatchewan and Manitoba: the Vermilion River, Favelle, and Ashville Formations. After deposition of this sequence, conditions in the geosyncline changed and a very thick accumulation of freshwater strata followed, interfingering eastward with brackish and freshwater deltaic deposits and marine strata of the shelf region (Fig. 51). In latest Cretaceous times the sea withdrew completely and continental deposition took place until the broad uplift along the Rocky Mountains belt marked the beginning of the Tertiary period.

The Tertiary uplift caused a rejuvenation of erosion, and thick deposits of freshwater clastics were spread far eastward, but much of those sediments was removed again by post-Tertiary erosion. The greatest thickness of these beds is preserved in the Alberta syncline, where Paskapoo Formation lies disconformably on the Edmonton Formation (Allan and Sanderson, 1945). Tertiary strata are also preserved on the southern chain of plateaux of Cypress Hills, Wood Mountain, Moose Mountain, and Turtle Mountain.

The sediment associations that resulted from these events are shown schematically in Figure 51. The post-Colorado sedimentation can be divided in the geosyncline association of Alberta and the intracratonic open-basin association of Saskatchewan and Manitoba. The transition of one association to the other is characterized by a west to east replacement of non-marine by marine sediments and of sand by shale.

The Cretaceous sandstones of the geosyncline association are typically grey to buff coloured subgreywackes, developed in massive crossbedded strata or in irregular agglomerations of lenses, interbedded with silty strata. The sorting is generally moderate, and particle sizes range from coarse sand to fine silt. Locally, the strata contain plant remains, and carbonaceous flakes are common.

Thick Upper Cretaceous sandstone sequences are best developed in Alberta, where they occur in the Eagle Formation (Milk River sandstone), in the sandstone members of the Belly River Formation, in the Upper Bearpaw Formation (Blood Reserve sandstone), and throughout the Edmonton Formation. Of these, only the sandstone members of the Belly River Formation and those of the Bearpaw Formation (Vermilion member, Ardkenneth sandstone, and Middle Beechy sandstone) extend into Saskatchewan. None of these reaches east of Last Mountain Lake.

Many of the Cretaceous sandstone strata in Alberta are associated with coal measures, particularly in the Edmonton Formation and in the upper members of the Belly River Formation. The Edmonton Formation contains fourteen coal seams, several of which are mined near Drumheller and Edmonton.

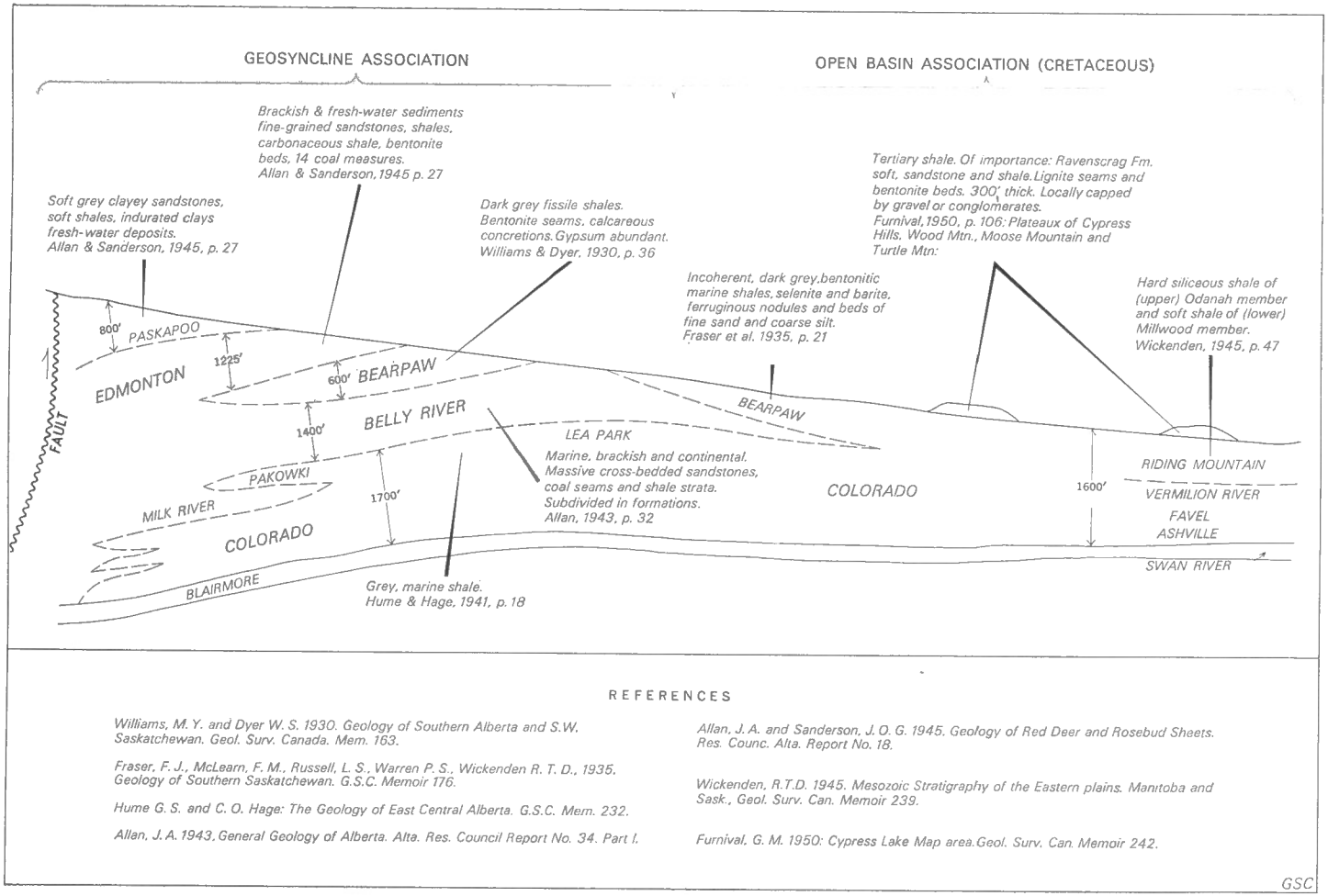


FIGURE 51. Schematic presentation of Cretaceous and Tertiary sediment associations of the Prairie Provinces (cross-sections after Shaw and Wells, 1954).

Another sand-rich formation is the Upper Cretaceous Eastend Formation in Saskatchewan and Alberta, which includes massive yellowish and yellowish green very fine sands and coarse silts. Laterally, the massive sands may pass into thin-bedded sand and shale. No carbonaceous materials or coal layers have been observed in the Eastend Formation. Fraser, *et al.* (1935, p. 26) summarized the nature of the Eastend Formation as follows:

> From west to east the beds thought to be coeval with the typical Eastend of Southern Saskatchewan are non-marine sands and shales in south-western and south-eastern Alberta, beds of much finer grain and transitional from marine to non-marine in southern Saskatchewan and possibly marine shales in Manitoba.

The thickest Tertiary sandstone strata occur in Alberta, in the Paskapoo Formation and its stratigraphic equivalents in southwestern Alberta, the Porcupine Hills and Willow Creek Formations. The age of these formations has been determined as Paleocene. The sandstone strata vary from soft, grey, clayey sandstones in the upper part of the formation to a coarse, more or less uncemented, sandstone in the lower part.

The Ravenscrag Formation in southern Saskatchewan, which is correlative with the Paskapoo Formation, also contains argillaceous sandstone beds, grey to greenish grey in colour, ranging in thickness from an inch to 10 feet, and interbedded with grey or black shale. Coal measures are common in the Ravenscrag Formation and there are mines throughout southern Saskatchewan (*see* Fraser, *et al.* 1935, pp. 63-88).

Little is known about the sandstones of the Turtle Mountain Formation in Manitoba. Wickenden (1945, p. 51) described the formation as follows:

> The outcrops observed consist mostly of fine-grained, white or yellowish sand and sandstone, with occasional fossil plant remains. Some shaly beds are present and also a few thin concretionary layers and beds of lignite.

Tertiary sandstones and conglomeratic beds of Eocene, Oligocene, and Miocene age are of local importance in southern Saskatchewan, where they occur in the Cypress Hills and Swift Current Formations.

Very little information is available about the permeability of these sandstones, although a large number of pumping tests has been carried out in these formations. As is explained in the next section, however, these pumping tests merely give an apparent permeability, not the true value. The indications are that the permeability of the sandstones ranges from 1 in./hr to 0.1 in./hr, with a gradual decrease in permeability from west to east (Milk River sandstone near Milk River, Alta. 0.62 in./hr; Ardkenneth sandstone near Riverhurst, Sask., 0.36 in./hr; Middle Beechy sandstone near Aylesbury, Sask., 0.14 in./hr).

As far as the shales are concerned the marine Cretaceous shales of the plains are characteristically bentonitic, soft, fissile, and dark grey or bluish grey in colour. Montmorillonite is their chief clay mineral. From west to east they change (Fig. 52) from soft carbonaceous shales (Edmonton Formation) to gypsiferous shales (Bearpaw Formation in Saskatchewan) to hard siliceous shales in Manitoba (Odanah Member of the Riding Mountain Formation). The outcrops of the gypsiferous shales are marked with selenite crystals and radiating balls of barite. Thin layers of calcareous

siltstone concretions or bands of clay-ironstone nodules can be seen in many outcrops. Layers of aragonite have been observed in association with the concretionary beds.

The permeability of the shales seems to be much more depth-dependent than the sandstone permeability. In the interval from zero to 150 feet typical shale permeabilities range from 0.02 to 0.002 in./hr. Kunkle (1962) found a "vertical permeability" of 0.0028 in./hr (0.035 g/d/ft^2) for 150 feet of Bearpaw shale overlying the Bulwark sandstone in central Alberta. Below 150 feet the permeability has been observed to decrease even further. For instance, the permeability of Bearpaw shale measured in piezometers near Bethune (Sask.) decreased from 0.07 in./hr at 100 feet to 0.00072 in./hr at 300 feet below surface.

Bentonite, which was mentioned several times in the previous paragraphs, occurs throughout the Cretaceous and Tertiary strata of the entire plains region (Ross, 1963). Well-known outcrops of this clay in Alberta are in the Edmonton Formation near Rosedale, Camrose, and Drumheller, and in the Bearpaw Formation near Walsh and Irvine. In Saskatchewan, bentonite is particularly common in the Ravenscrag Formation, in the Battle Formation in southwestern Saskatchewan, and in the Vermilion Formation of eastern Saskatchewan. In Manitoba, bentonite is common in the Vermilion Formation and Riding Mountain Formation, as can be seen at various places along the Manitoba escarpment.

Swelling is the most characteristic property of bentonite with respect to water, while chemically the sodium ions that are attached to the crystal framework of montmorillonite can be exchanged for calcium and magnesium.

Sedimentary Rocks of the Eastern Plains

In Manitoba, the Cretaceous strata end abruptly at the Manitoba escarpment (Fig. 52). East of the escarpment extends the Manitoba lowland, which is underlain by flat-lying Jurassic and Palaeozoic rocks, ranging in age from Ordovician to Devonian.

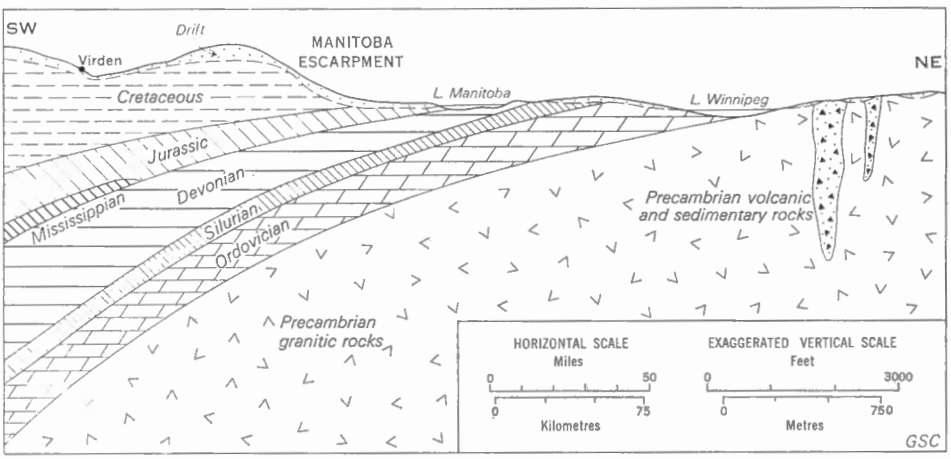


FIGURE 52. Geological cross-section of southern Manitoba (*from* Davis, *et al.*, 1962).

The Ordovician comprises nearly 1,000 feet of strata, including 50 to 100 feet of basal white quartz sandstone (Winnipeg Formation), overlain by 550 feet of limestone and dolomite (Red River Formation), in turn overlain by red shales and variegated argillaceous limestones and dolomites of the Stony Mountain Formation. The Upper Ordovician is represented by 50 to 60 feet of grey dolomite, the Stonewall Formation.

The Silurian in Manitoba is made up of 400 feet of finely textured dolomites (Interlake Group) overlain conformably by Middle Devonian limestones and dolomites, reaching a thickness of 1,400 feet in the subsurface of southwestern Manitoba (Fig. 52). Outcrops of Middle Devonian strata, belonging to the Elk Point Group and Winnipegosis Formation, occur in the region between Lake Winnipegosis and Lake Manitoba.

The Winnipegosis Formation contains hard, dense, biohermal and non-biohermal dolomites and an interreef facies of bedded dolomite and thin shaly bands. The Devonian strata extend into the northeastern plains as a narrow band along the margin of the shield.

No surface exposures of Jurassic rocks are known in the Manitoba lowland. The subsurface section has been divided in ascending order into Amaranth, Reston, Melita, and Waskada Formations, with a total thickness of 1,000 feet. The sediments range from anhydrite, gypsum, and dolomite beds in the Amaranth Formation to argillaceous limestone and varicoloured shales in the upper part of the section.

Hydrogeological Bedrock Regions of the Western Canada Sedimentary Basin

The regional distribution of the materials discussed in the two previous sections is shown in Figure 53 (*in pocket*). The aquifer properties of these hydrogeological units are discussed in the next section.

Bedrock Strata as Aquifers

The irregular Cretaceous sedimentation patterns of the transitional environment in the geosyncline area precluded the deposition of widespread thick blanket sandstones. As a result, none of the Cretaceous formations contains extensive uniform aquifers. Nevertheless, many of the agglomerations of irregular sandstone lenses and many of the strata of limited areal extent have sufficient permeability to be functional aquifers.

Regionally, both the permeability and the cumulative thickness of the sandstone beds in the stratigraphic column decrease from west to east, and consequently bedrock aquifers become poorer from west to east. This tendency is sufficiently demonstrated by the data that are presented in Table XIV.

The cumulative thickness of the sandstone layers penetrated by one well bears some relationship to its water-yielding characteristics, but the best wells are completed where one or more thick sandstone beds are encountered, although their cumulative thickness may not be great. Since the term "transmissibility" has no meaning under

such circumstances, the Research Council of Alberta introduced the term "apparent transmissibility" (Farvolden, 1961) as a measure of the permeability of the "aggregate aquifer." The value is obtained from a simple pumping test, by applying the modified non-equilibrium equation to a straight-line plot of water-level drawdown against the logarithm of the duration of the test in minutes. It should be pointed out that these values incorporate well loss effects and are therefore generally too low.

Table XIV

Transmissibility Values (g/d/ft) in Bedrock Aquifers of Interior Plains Region

Number[1]	Formation and member	Area	Transmissibility (range and average)	Reference
1	Wapiti	Beaverlodge	300–7700 (2475)	Jones, 1960
2	Paskapoo–Edmonton	Swan Hills	17–2640[2] (695)	Jones, 1962
3	Paskapoo	Pembina	300–1000[2]	Farvolden, 1961
4	Paskapoo	Blindman R.	1300–7200 (4268)	Nielsen, 1963
5	Paskapoo	Calgary	150–250[2]	Meyboom, 1961
6	Edmonton	Edmonton	55–2400 (818)	Lennox, 1964
7	Edmonton	Stettler	3000	Lennox, 1964
8	Belly River Pale and var. beds	Andrew	6–53 (25)	LeBreton, 1963a
9	Belly River Pale and var. beds	Vegreville	100	LeBreton, 1963c
10	Belly River Birchlake	SE Alberta	2600	LeBreton, 1963c
11	Belly River Ribstone Creek	Lloydminster	900–4700 (2700)	LeBreton, 1963b
12	Eagle Milk River ss.	S. Alberta	8–3000 (370)	Meyboom, 1960
13	Bearpaw Bulwark ss.	Oyen	500	Kunkle, 1962
14	Bearpaw Ardkenneth ss.	Outlook	1300	GSC files
15	Bearpaw Ardkenneth ss.	Riverhurst	260	GSC files
16	Bearpaw Beechy ss.	Bethune	203	GSC files
17	Blairmore	Regina	880	Lissey, 1962
18	Eastend	Shaunavon	5640	Big Indian Drilling Co.
19	Riding Mountain Odanah	Strathclair	4950	GSC files
20	Red River	St. Pierre	6422	Charron, 1964
21	Red River	Winnipeg	30,000	Hobson, *et al.*, 1964

[1]Location of areas is shown by numbers on the map of the hydrogeological bedrock regions (Fig. 53).
[2]"Apparent transmissibility"; for explanation *see text.*

The bedrock aquifers of the prairie region are sufficient everywhere to meet domestic water requirements, but sandstone aquifers that can yield 50 gpm or more are limited to the Paskapoo Formation, parts of the Edmonton Formation, the Birchlake and Ribstone Creek Members of the Belly River Formation, and perhaps some parts of the Eastend and Ravenscrag Formations. With respect to municipal water supplies, only the Paskapoo Formation and the Edmonton Formation are tapped for municipalities of up to 2,500 inhabitants. The corresponding production of 125,000 gpd (85 gpm) appears to be, on the whole, the ultimate safe yield of these formations.

As far as their hydrologic behaviour is concerned, most bedrock aquifers appear to be confined; that is they are not intersected by the phreatic surface. The determination of their physical properties can be done successfully with the Theis non-equilibrium equation or Jacob's modification thereof. Closer examination reveals that under conditions of continuous pumping many of the sandstone aquifers receive leakage from the overlying silty shale, and for this reason longer pumping tests and application of the leaky artesian equations may give better approximations of the actual conditions than the non-leaky formulae.

Areas with "flowing artesian wells" are common throughout western Canada. Their occurrence is determined by the combination of permeable material and an increase in total head with depth, a condition generally found in discharge areas and in the transmission areas of stratified media. Recharge to such artesian aquifers is derived from the overlying "confining beds," and the extent of the recharge areas is governed entirely by the elevation of the overlying phreatic surface.

Although Figure 53 suggests that good aquifers can be found only east of Last Mountain Lake, and particularly in Alberta, the siliceous shale member of the Riding Mountain Formation (Odanah beds) is an important aquifer for domestic supplies in rural Manitoba. The yield from the hard fractured shale, which is seldom more than 100 feet thick, varies from 5 to 10 gpm but can be as high as 50 gpm.

The aquifers in the Palaeozoic limestone have high transmissibilities (Table XIV), but high permeability is believed to be limited to a relatively thin solution zone parallel to the surface (Hobson, Scott, and van Everdingen, 1964). The most serious drawback to the exploitation for human consumption of Palaeozoic aquifers in Manitoba is that many contain saline water (Charron, 1961, 1964).

Hydrochemistry

Hydrochemistry of Surficial Deposits

Groundwater acquires its principal chemical composition during infiltration in the soil. In the soil zone, where the net amount of downward flow is the balance of infiltration and evaporation, two chemical processes take place: dissolution and precipitation. According to Schoeller (1959), the principal process in temperate climates is dissolution. In semi-arid areas, however, precipitation (and hence concentration) of salts in the soil acquires more importance, owing to the preponderance of capillary rise over rainfall. Schoeller commented that concentration is particularly noticeable after light rainfall followed by strong evaporation. In such areas, water

that finally reaches the zone of saturation after a period of intense infiltration will be more mineralized than water that infiltrates in temperate areas. Thus, rainfall and temperature are major factors in establishing a climatological macrozonation of groundwater chemistry.

This macrozonation is best expressed by the concentrations of the most soluble salts, principally $CaSO_4$ and NaCl. The effect is less noticeable for $CaCO_3$ whose solubility is determined by the amount of free CO_2, rather than by the amount of rainfall. Applying these general principles to the prairie environment, one may expect to find a hydrochemical reflection of the low-rainfall areas in southern Saskatchewan and Alberta (less than 16 inches), as compared to the regions of higher rainfall in western Alberta and east-central Manitoba (more than 18 inches).

This was demonstrated by LeBreton and Jones (1962, Fig. 13), who showed that groundwater in the till of western Alberta is low in total dissolved solids (less than 800 ppm) and contains predominantly Ca-Mg bicarbonates. Eastward, the average amount of total dissolved solids increases to 2,500 ppm in east-central Alberta and central Saskatchewan (Christiansen, 1961), and sulphates become the predominant constituents—both sodium sulphate and calcium-magnesium sulphate. Farther east, toward southwestern Manitoba, the average amount of total dissolved solids diminishes to 2,000 ppm (Halstead, 1959), and to 1,100 ppm in southern Manitoba (Charron, 1961). Calcium-magnesium sulphates are the predominant dissolved salts in both areas. The eastern boundary, between shallow sulphate water and bicarbonate water, was found by Charron (1964) in the Fannystelle area of central Manitoba. There, the average amount of total dissolved solids is 912 ppm, and calcium-magnesium bicarbonate dominates over calcium-magnesium sulphate in the proportion of about 2:1.

On the whole, groundwater in surficial deposits in areas with a total annual precipitation of more than 18 inches appears to be moderately mineralized (less than 1,000 ppm) and of the calcium-magnesium bicarbonate type. In areas with less than 16 inches annual precipitation groundwater in drift contains 1,000-2,500 ppm total dissolved solids and is of the calcium-magnesium sulphate or sodium sulphate type. The chemistry of shallow groundwater on the prairies can be interpreted so well in terms of Schoeller's macrozonation largely because of the uniformity of the surficial deposits. However, this simple hydrochemical zonation is modified once the drift water reaches the less uniform bedrock strata.

Hydrochemistry of Shallow Bedrock Aquifers (Upper Cretaceous and Younger)

Figure 55 (*in pocket*) shows the hydrochemical map of the shallow bedrock aquifers on the prairies. The map has been constructed by superposition of 4 base-maps, each one showing the geographic variation of one of the following ratios:

$$r\frac{Na}{Ca+Mg};\quad r\frac{Ca}{Mg};\quad r\frac{Cl}{HCO_3+SO_4};\quad \text{and}\quad r\frac{HCO_3}{SO_4}.$$

The ratios were computed for 229 complete analyses of samples taken from a depth of less than 400 feet. Groundwater to that depth contains between 1,000 and 3,000

ppm total dissolved solids. Below 400 feet the degree of mineralization increases rapidly in most areas.

The hydrochemical map shows six categories of groundwater: calcium bicarbonate water, magnesium bicarbonate water, sodium bicarbonate water, calcium-magnesium sulphate water, sodium sulphate water, and sodium chloride water. Most of the hydrochemical zonation of bedrock waters can be explained by chemical alterations of the sulphate or bicarbonate drift waters brought about by typical bedrock materials. These alterations are related to the presence of lignite, coal, methane, bentonite, and selenite.

Bedrock groundwater in most of southern Alberta is of the sodium bicarbonate type. The transformation of shallow calcium-magnesium sulphate water or sodium sulphate water to bicarbonate water is believed to be brought about by bacterial sulphate reduction (genus *Sporovibrio*) in the presence of organic hydrocarbons. The process is of the type described by Hem (1959, p. 223):

$$SO_4 + CH_4 \xrightarrow{\text{bacteria}} HS^- + H_2O + HCO_3^-.$$

Another process involved in the sulphate-bicarbonate transformation is related to the much larger amount of CO_2 in coal measures than that present in the air of the topsoil. Consequently, the HCO_3 content of groundwater in coal seams is much greater than the normal values of 5–10 milliequivalents/litre. This is demonstrated in Figure 54, which shows the HCO_3 content of 90 random samples of bedrock waters in the Prairie Provinces. Forty-seven per cent of these have a normal HCO_3 content between 5 and 10 epm, 23 per cent have below-normal values, and 30 per cent have above-normal values. Leaving the below-normal concentrations for a later discussion, it can be said that the above-normal concentrations were determined in samples from the coal-bearing Edmonton and Ravenscrag Formations.

The change from predominantly calcium and magnesium in the drift to sodium in the bedrock waters is related to a base exchange with the bentonite. The process is generally given by

$$\underset{\text{solution}}{Ca[(HCO_3)_2 \cdot SO_4]} + \underset{\text{solid}}{Na_2} \rightleftarrows \underset{\text{solution}}{Na_2[(HCO_3)_2 \cdot SO_4]} + \underset{\text{solid}}{Ca}.$$

The base exchange in parts of the Edmonton Formation has led to ratios of sodium over calcium plus magnesium $\left(\frac{Na}{Ca+Mg}\right)$ of more than 50.

The combined effects of these three processes determine the very alkaline nature of bedrock waters in all formations that contain coal, lignite, and bentonite; these are, particularly, the Edmonton Formation and its stratigraphic equivalents, the Belly River Formation and, in southern Saskatchewan, the Ravenscrag Formation. The combination of only base exchange and sulphate reduction may take place in the presence of bentonite and methane, as was postulated by Meyboom (1960) for the Milk River sandstone in southern Alberta.

In Saskatchewan, the absence of Cretaceous coal and the presence of gypsiferous shales tend to emphasize the sulphate character of the drift water, and most of the

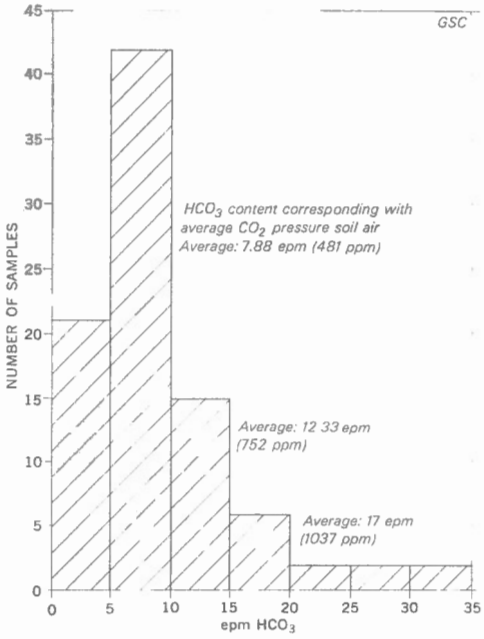


FIGURE 54.
Bicarbonate content of 90 random water samples from shallow bedrock aquifers.

Cretaceous aquifers in Saskatchewan contain sulphate water. The hydrochemical map shows a distinct differentiation of calcium-magnesium sulphate water around a central area of sodium sulphate water. Some of the below-average HCO_3 contents in Figure 55 (*in pocket*) are from samples in this central zone, indicating that the dissolution of $CaSO_4$, which lowers the HCO_3 concentration, is an active process in the marine shales of this area.

The influence of higher rainfall in the Cypress Hills area is shown by the magnesium bicarbonate subzone in that region. The alteration to sodium bicarbonate water in the Shaunavon area has to be related to base exchange in the bentonite beds of the Ravenscrag Formation. Parsons (1964) also reported sodium bicarbonate water from the Eastend and Frenchman Formations of this area.

Sodium bicarbonate waters reappear in the Riding Mountain Formation of southern and south-central Manitoba. There, the bicarbonate contents are normal, and the transformation of the shallow magnesium-calcium bicarbonate water to the sodium bicarbonate water may be caused by base exchange in the bentonitic shale.

The occurrence of sodium chloride water at shallow depth is uncommon on the Prairies. Such water is found only in the Carnduff artesian area (south Saskatchewan) and in the limestone aquifers of the Manitoba lowland. Here, the bedrock aquifers contain an average of 3,000 ppm total dissolved solids, half of which are chlorides (Charron, 1961, 1964). The appearance of sodium chloride water in the Manitoba lowland may be related to very saline water in the deeper Palaeozoic formations that are believed to ascend into the Red River valley. Below-normal HCO_3 concentrations are common in the sodium chloride subzone.

Hydrochemistry of Deep Bedrock Aquifers (Lower Cretaceous and Older)

The following résumé of the hydrochemistry of deep bedrock aquifers has been prepared by R. O. van Everdingen from analyses supplied by the Mines Branch of the Manitoba Department of Mines and Mineral Resources (Winnipeg, Manitoba), by the Petroleum and Natural Gas Branch of the Saskatchewan Department of Mineral Resources (Regina, Saskatchewan), and by the Alberta Oil and Gas Conservation Board (Calgary, Alberta).

Chemical analyses of formation waters, made by or for the major petroleum companies and the Provincial Oil and Gas Boards, are available for the majority of the deep aquifers of the western Canada Sedimentary Basin. The study of this information has just been started, and so far only major trends can be indicated.

Of all the formation waters studied, the highest concentrations of total dissolved solids (up to 300,000 ppm) are found associated with the deeper parts of the original basins along the Foothills Belt (Edmonton-Red Deer area, Lloydminster area, Swan Hills area) and in the Williston Basin (Frobisher area, southeast Saskatchewan). The lowest concentrations, often below 5,000 ppm, occur along the subcrops near the Foothills Belt and over the Sweet Grass Arch in southern Alberta and southwestern Saskatchewan.

In the areas of high concentration the total dissolved solids increase rapidly with depth. For instance, the Midale and Frobisher analyses (Table XV, area 1) show an increase from 107,000 ppm to 256,000 ppm over a vertical distance of 75 to 150 feet. In general, the high values of total dissolved solids are probably a function of depth below surface and distance from the outcrop area, in combination with the occurrence of halite deposits; e.g., in the Devonian Prairie Formation.

For waters with total dissolved solids of 10,000 ppm or less, the following ionic relations were found in the following sequence of frequency:

1. $\left.\begin{matrix} rCl^{1} > rHCO_3 > rSO_4 \\ rCl > rSO_4 > rHCO_3 \end{matrix}\right\}$ $rNa > rCa > rMg$
2. $rSO_4 > rCl > rHCO_3$ $rCa > rNa > rMg$
3. $rHCO_3 > rCl > rSO_4$.

For waters with a total mineralization of more than 10,000 ppm the general characteristics are:

4. $rCl > rSO_4 > rHCO_3$
   or $rCl > rHCO_3 > rSO_4$

with Cl making up 40–49.9 per cent of the equivalent parts per million, and

$$rNa > rCa \geqslant rMg$$

with Na making up 27–48.9 per cent of the equivalent parts per million.

The characteristics of the formation waters presented in Table XV can be summarized as follows:

1. *Frobisher area, Saskatchewan*

   High to very high total dissolved solids; sodium chloride waters. In the Cretaceous formations some bicarbonate is present and the waters have relatively low total

[1] r denotes equivalent parts per million.

Table XV

Hydrochemistry of Deep Bedrock Aquifers of Alberta, Saskatchewan and Manitoba

Formation	Na	Ca	Mg	Cl	SO_4	HCO_3	CO_3	T.D.S.[1]
1. FROBISHER AREA, SASKATCHEWAN								
Cretaceous								
Viking	48.57	0.94	0.49	45.17	0.19	4.54		19,700
U. Blairmore	48.92	0.88	0.20	40.66	0.39	7.45	1.50	7,000
L. Blairmore	48.82	0.78	0.40	43.08	4.80	1.87	0.25	20,000
Jurassic	48.39	1.01	0.60	40.09	8.85	0.98	0.08	25,000
Mississippian								
Midale	45.22	3.81	0.97	47.26	2.64	0.10		107,000
Frobisher	45.98	2.70	1.32	49.56	0.40	0.04		256,000
Lodgepole	44.98	3.75	1.27	49.41	0.42	0.17		222,000
Devonian								
Nisku	39.73	8.56	1.71	49.79	0.18	0.03		296,000
Duperow	35.14	12.22	2.64	49.75	0.18	0.07		270,000
Souris R.	33.30	13.78	2.92	49.86	0.11	0.03		271,000
Winnipegosis	42.70	3.40	3.90	49.91	0.05	0.04		302,600
Silurian	46.78	2.52	0.70	49.95	0.51	0.04		261,000
Ordovician								
Red River	36.60	11.28	2.12	49.88	0.10	0.02		311,000
Winnipeg	43.07	5.36	1.57	49.83	0.16	0.01		297,000
Cambrian	37.31	11.51	1.18	49.92	0.07	0.01		294,000

Formation	Na	Ca	Mg	Cl	SO_4	HCO_3	CO_3	T.D.S.[1]
2. GRAVELBOURG-WOOD MOUNTAIN AREA								
Cretaceous								
Viking	47.20	1.80	1.00	48.00	0.40	1.60		8,000
U. Blairmore	48.50	0.90	0.60	36.90	1.50	9.80	1.80	3,500
L. Blairmore	46.80	1.40	1.80	20.10	17.60	10.00	2.30	3,000
Jurassic	47.90	1.30	1.80	34.00	9.10	5.70	1.20	3,600
Mississippian	16.30	22.80	10.90	3.30	43.30	3.60		3,200
Devonian								
Nisku	41.90	4.10	4.00	8.10	37.70	4.20		4,000
	14.10	23.80	12.00	10.90	37.20	1.90		4,700
Duperow	34.30	10.10	5.60	12.80	32.10	5.10		6,300
	9.20	27.90	12.90	3.50	43.20	3.30		3,000
Souris R.	38.60	7.20	4.20	42.50	7.00	0.50		21,800
	25.20	13.60	11.20	11.20	35.40	3.40		2,900
Dawson Bay	29.30	16.80	3.90	26.60	21.50	1.90		9,500
	16.80	24.00	9.10	17.20	31.00	1.80		4,000
Winnipegosis	27.80	16.70	5.50	13.30	33.30	3.40		5,800
Silurian	31.20	14.30	4.50	26.60	21.90	1.50		9,500
Ordovician								
Red River	42.60	5.70	1.70	45.80	4.00	0.20		73,100
Winnipeg	22.40	19.40	8.20	10.00	37.60	2.40		3,900
Cambrian	40.60	6.40	3.00	48.40	1.50	0.10		107,000

3. SOUTHERN PLAINS, ALBERTA: (A) TABER; (B) PRINCESS AREA

Cretaceous								
(A) Milk River type D	48.00	1.50	0.50	44.00	0.50	5.00	0.50	5,000
(A) Viking	49.00	0.70	0.30	44.00	0.50	5.00	0.50	10,000
(B) Viking	48.50	1.00	0.50	48.50	tr	1.50		25,000
(A) Taber	48.50	0.70	0.80	31.50	0.20	18.30		8,000
(B) L. Cretaceous	49.00	0.50	0.50	39.00	0.50	8.50	2.00	30,000
(A) Ellis/Miss	48.00	1.00	1.00	15.00	1.00	34.00		7,000
Mississippian								
(B) Madison	49.00	0.50	0.50	37.00	0.50	12.50		25,000
(B) Banff	42.00	6.00	2.00	36.00	13.00	1.00		20,000
Devonian								
(B) Three Forks	44.00	3.00	3.00	25.00	17.00	8.00		20,000
(B) Jefferson 1st por.	38.00	9.00	3.00	10.00	31.00	9.00		10,000
(B) Jefferson 2nd por.	42.00	6.00	2.00	38.00	8.00	4.00		20,000
Siluro-Cambrian								
(B)	27.00	21.00	2.00	50.00	tr	tr		300,000

[1]Total dissolved solids are reported in parts per million, whereas the cations and anions are reported in per cent milliequivalents. Analyses given are averages of two or more analyses.

4. REDWATER AREA, CENTRAL ALBERTA

Cretaceous							
Belly River	44.40	4.80	0.80	49.86	0.08	0.06	17,000
Viking	44.50	3.20	2.30	49.24	0.06	0.70	50,000
Basal Quartz	43.90	3.80	2.30	49.17	0.03	0.80	71,000
Devonian							
Wabamun	43.80	3.90	2.30	49.25	0.02	0.73	72,000
Winterburn	41.80	5.40	2.80	49.89	0.01	0.10	82,500
Cooking Lake	40.10	5.50	4.40	49.08	0.79	0.13	93,400
Leduc	37.90	6.40	5.70	49.30	0.50	0.20	101,700
Beaverhill Lake	34.60	9.10	6.30	48.60	1.20	0.20	112,600
Cambrian	47.40	1.40	1.20	49.50	0.40	0.10	295,000

5. NORTHERN ALBERTA, SWAN HILLS AREA

Cretaceous							
Viking	46.20	2.00	1.80	49.60	0.02	0.38	41,000
Gething	46.60	1.80	1.60	49.40	0.20	0.40	62,000
Jurassic	46.40	2.80	0.80	49.60		0.40	116,000
Mississippian							
Rundle	43.10	5.50	1.40	49.00	0.20	0.80	65,000
Banff	46.10	3.10	0.80	49.50	0.20	0.30	129,000
Devonian							
Winterburn	34.70	12.20	3.10	49.50	0.10	0.40	174,000
Cooking Lake	35.80	12.20	2.00	49.70	0.10	0.20	204,000
Beaverhill Lake	45.10	3.50	1.40	49.70	0.20	0.10	195,000
Gilwood	33.90	13.30	2.80	49.85	0.05	0.10	229,000
Elk Point	34.30	12.00	3.70	49.92	0.03	0.05	293,000
Cambrian	33.70	13.10	3.30	49.96	0.01	0.03	250,000

dissolved solids. In the Lower Cretaceous to Mississippian interval the sulphate content may be up to 8.9 per cent. In the Devonian and lower, the calcium content may be as high as 14 per cent.

2. *Gravelbourg–Wood Mountain area, Saskatchewan*
   Low total dissolved solids; the waters have an irregular composition.
3. *Taber and Princess areas, Alberta*
   Low to fairly high total dissolved solids; predominantly sodium chloride waters, although some sulphate waters are present in the Lower Mississippian and Devonian. In the Silurian and Cambrian, the calcium and magnesium contents are as high as 21 per cent.
4. *Redwater area, Alberta*
   Medium to high total dissolved solids; predominantly sodium chloride water. The calcium content is highest in the Devonian, where it reaches 9 per cent.
5. *Swan Hills area, Alberta*
   High to very high total dissolved solids; sodium chloride waters. In the pre-Mississippian formations the calcium content may be as high as 14 per cent.

Brine Springs along the Precambrian Shield

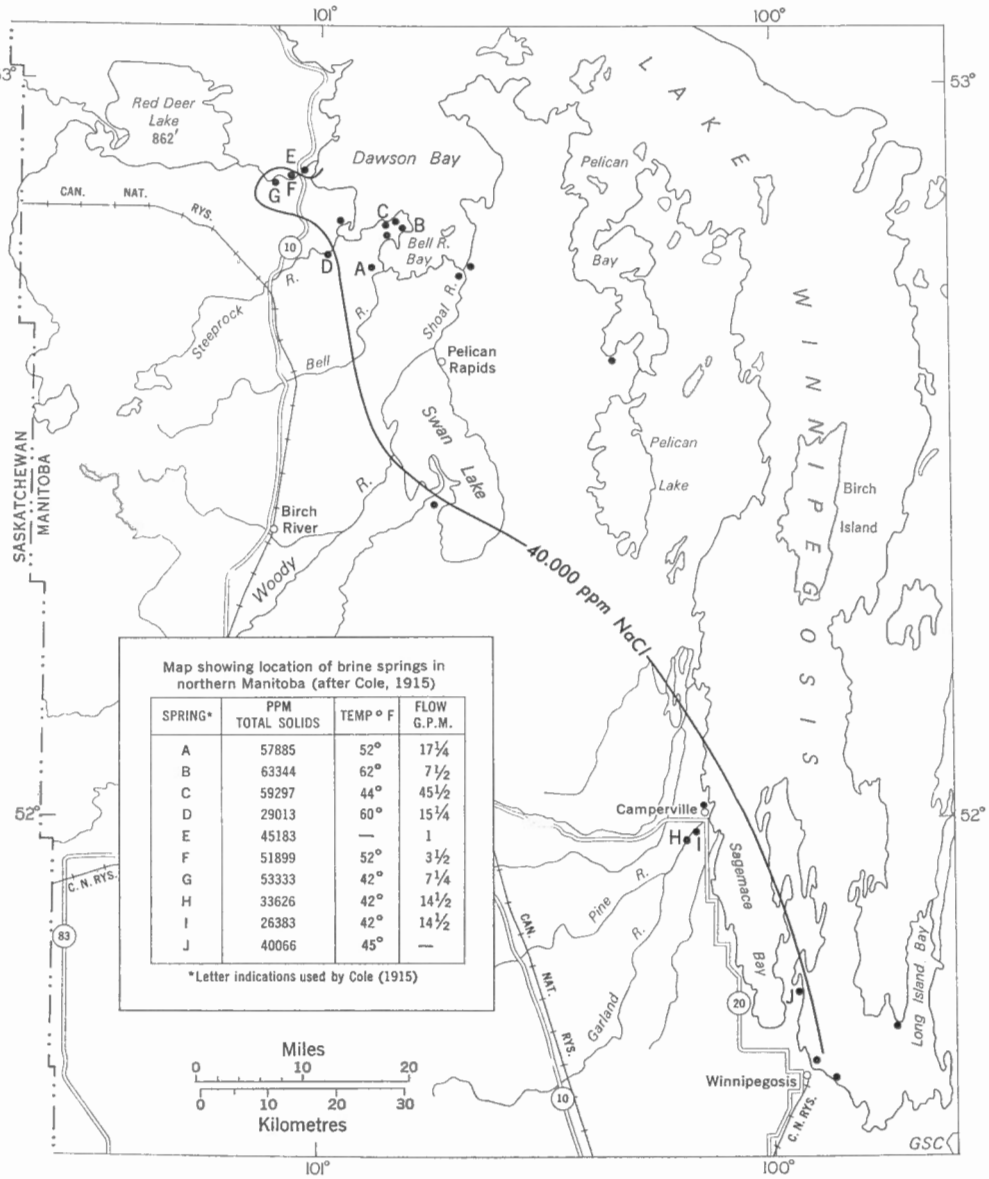
The hydrochemistry of the regions underlain by Palaeozoic and Jurassic rocks stands apart from the sulphate-dominated Interior Plains. Numerous brine springs, containing more than 25,000 ppm total solids, issue from the Palaeozoic strata along Lake Winnipegosis and Lake Manitoba, and diffuse seepage of saline groundwater into these lakes is evident from the composition of lake waters, which contain up to 600 ppm chlorides (Thomas, 1959).

The Palaeozoic springs (Fig. 56) fall into two groups, separated by the 40,000 ppm isochlor, whose direction coincides with the strike of the strata. The division appears to be related to the stratigraphic boundary between the light-grey upper Manitoban limestone (less than 40,000 ppm NaCl) and the lower Manitoban limestone plus the Winnipegosis dolomite (more than 40,000 ppm NaCl). Thus the entire Upper Devonian section contributes to the supply of brine water.

Temperature measurements along the western shore of Dawson Bay (Lake Winnipegosis), carried out by Cole (1915), suggest that the brines originate in an even larger stratigraphic interval. There is a ten-degree difference in average brine temperatures east and west of the 40,000 ppm isochlor, which at this location would indicate a difference in depth of origin of about 500 feet.[1] This difference in depth of origin coincides reasonably well with the thickness of the *entire* Devonian section underneath the Manitoba escarpment, as it was described by Wickenden (1945, Fig. 2).

Chemically, the brines fall within Palmer's group 4 (90% primary salinity caused by NaCl, 10% secondary salinity due to chlorides and sulphates of Ca and Mg). They contain up to 96 ppm bromide, and the equivalent ppm bromide/iodide for

[1]The mean annual temperature of the Winnipegosis district is 32°F, whereas the average brine temperatures vary from 42° to 52°F. With a geothermal gradient of 0.02°F/ft, the water of 42°F can be estimated to come from a depth of 500 feet, and the water of 52°F from 1,000 feet.



Map showing location of brine springs in northern Manitoba (after Cole, 1915)

SPRING*	PPM TOTAL SOLIDS	TEMP° F	FLOW G.P.M.
A	57885	52°	17¼
B	63344	62°	7½
C	59297	44°	45½
D	29013	60°	15¼
E	45183	—	1
F	51899	52°	3½
G	53333	42°	7¼
H	33626	42°	14½
I	26383	42°	14½
J	40066	45°	—

*Letter indications used by Cole (1915)

FIGURE 56. Location and degree of mineralization of brine springs in northern Manitoba (after Cole, 1915).

water from the lower Upper Devonian in the Neepawa salt well is approximately 1 (96 ppm Br and 15 ppm I).

The few springs that issue from the Jurassic rocks near Lake Manitoba have less total dissolved solids (less than 10,000 ppm), but the dominant ions are also sodium and chloride. The Jurassic brines, also, belong to Palmer's class 4.

Brine springs occur north of the Interior Plains region and are mentioned here because of their similarity. In the Slave River district, a number of brine springs rise from the Silurian rocks at the base of the escarpment west of Fort Smith. Salt was gathered from two of these localities by the Hudson Bay Company (Camsell and Malcolm, 1921) and the Roman Catholic Mission. The flow of the springs, which issue from the Fitzgerald Formation, is not more than 4 gpm. The brines contain 260,000 ppm total solids, which is of the same order as the amount of total dissolved solids of the waters in the deep aquifers of the Swan Hills area listed in Table III. Ninety-eight per cent of the constituents is sodium chloride.

The best-known brine spring along the Mackenzie River is La Saline spring, 28 miles downstream from the junction with Clearwater River. The brine contains 70,000 ppm total dissolved solids, of which 90 per cent are sodium chlorides. In his recent description of the geology of that area, Carrigy (1959) postulated that the saline waters come from the Devonian rocks that underlie the river valley. He suggested that the origin of the sodium chloride has to be sought in the halite beds of the Prairie Evaporite Formation (Elk Point Group), from which considerable solution has taken place.

The Water Balance

The order of magnitude of the amount of annual groundwater replenishment has been determined from hydrometric records using Wundt's method of analysis (Wundt, 1958). The method, which is designed to calculate the amount of recharge, is based on the assumption that most groundwater discharge from an area takes place in the rivers, and that in a steady-state system the amount of discharge is equal to the amount of recharge. As hydrometric observations do not account for underflow, much of which will be groundwater discharge, nor for evapotranspiration by riparian vegetation, which may reduce low-flow by as much as 50 per cent, Wundt's analysis gives merely an index of groundwater recharge, from which an absolute value has to be estimated.

According to Wundt, the monthly minimum streamflow (MN_q) is taken as a measure of groundwater run-off, and hence the annual mean monthly minimum ($\frac{1}{12}\sum MN_q$) is taken as a measure of annual groundwater run-off. The ten-year average MN_q, from 1950 to 1960, expressed in gallons per minute per square mile (g/m/sq. mi.) has been computed for 98 permanent hydrometric stations across the Prairies. The individual values showed considerable variation from one drainage basin to the next, but four broad regions of similar low-flow conditions could be outlined (Fig. 56) by the contours indicating indices of 5, 50, and 300 g/m/sq. mi.

Figure 55 shows that the base-flow index on the plains is everywhere less than 50 g/m/sq. mi., and in most areas less than 5 g/m/sq. mi. Expressed in inches of water and paying due regard to the principle of continuity, these values indicate amounts of natural groundwater replenishment ranging from 1.8 inches per year (in./yr) to 0.18 in./yr, the latter value being the index for recharge conditions over most of the plains.

If minimum streamflow is taken as a measure of groundwater discharge, one has to distinguish between the three groundwater components that make up baseflow: bank storage, contact springs, and artesian leakage. Hydrograph studies of prairie rivers (Meyboom, 1961; LeBreton, 1963) indicated clearly that bank storage may make up the bulk of base flow. Meyboom (1961) estimated that 73 per cent of the low flow of the Elbow River near Calgary was derived from bank storage, and LeBreton calculated that 78 per cent of the base flow in the Battle River came from that source.

The hydrochemistry of low-flow waters supports the argument that the bulk of low flow is derived from bank storage. As shown on the inset of Figure 55, the main type of low-flow water on the Prairies is calcium-magnesium bicarbonate water. In the semi-arid plains this is a typical indication of an environment where the concentration-infiltration effects of low rainfall and high evaporation are not noticeable; that is, a condition which may be expected to exist where floodwaters infiltrate into the river banks. The only areas where groundwater chemistry is reflected by the hydrochemistry of low-flow water is in the internal drainage basin of Old Wives Lake and in the headwater regions of the Qu'Appelle and Souris Rivers.

If as much as 70 per cent of the total base flow is derived from bank storage, the index of groundwater discharge from surficial and bedrock aquifers, in excess of underflow and evapotranspiration, has to be within the range of 0.54 in./yr to less than 0.05 in./yr. Farvolden (1963c) determined discharge from a till "aquifer" near Edmonton at 0.37 in./yr, excluding evapotranspiration. Nielsen (1963) calculated an average discharge from surficial and Paskapoo aquifers in the Blindman River basin of 0.32 in./yr. LeBreton (1963) concluded from stream hydrographs that discharge from surficial and bedrock aquifers in eastern Alberta amounted to 0.05 in./yr. These determinations are incidental values, but in the absence of more elaborate studies they are believed to be worth presenting in order to indicate what rates of groundwater discharge in excess of underflow and evapotranspiration have actually been observed on the Canadian Prairies.

If we introduce the factor 2 to account for underflow and evapotranspiration, the estimated total groundwater recharge, including recharge to bank storage, amounts to 3.6 in. to 0.36 in./yr, whereas recharge to surficial and bedrock aquifers alone would amount to 30 per cent thereof; i.e., 1.08 in. to 0.1 in./yr. Within the range of 16 to 12 inches of total annual precipitation, total recharge is thus estimated to lie within the range of 22 per cent to 3 per cent of the annual precipitation, and recharge to surficial and bedrock aquifers alone varies from 7 per cent to less than 1 per cent of the total annual precipitation.

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Chapter VII

CORDILLERAN HYDROGEOLOGICAL REGION

by E. C. Halstead

Introduction

The Cordilleran region includes a massive mountainous complex that extends from the Interior plains to the Pacific Ocean and from the 49th parallel to the Yukon. It is some 500 miles wide and extends almost 1,000 miles southeast from the southern limit of discontinuous permafrost to the International Boundary. It is a forested area with mountains projecting above timberline except in the south where lower drier hills support a vegetative cover of grasses and sparse woodlands.

Precipitation on the high peaks may be retained as glaciers and ice-fields whereas elsewhere it is shed as run-off, although weathered and fractured bedrock absorbs considerable water and transmits it for short distances and permeable overburden provides natural facilities for its storage. Sufficient data are not available, especially in areas of high altitudes and in the northern part, to accurately calculate the total annual precipitation on the Cordilleran hydrogeological region but it has been estimated at 800 million acre feet[1]. The run-off for the same area amounts to 650 million acre feet, so therefore more than 150 million acre feet are locked up as glacier ice, evaporated from surface bodies of water, lost through evapotranspiration, or stored in fractures and pores in the bedrock and overburden.

Physical Features

The Cordilleran hydrogeological region includes three major physiographical subdivisions, each of which is called a system. The three systems run the length of the Cordillera in a northwest direction and their topography controls to a large extent the behaviour of the Pacific and Arctic air masses that are responsible for the climate of the region.

The Western system includes the St. Elias Mountains, the ranges of the Queen Charlotte Islands and of Vancouver Island, and the Coast and Cascade Mountains. The St. Elias and Coast Mountains, which occupy most of the Western system, reach elevations of more than 15,000 and 13,000 feet, respectively, and nourish alpine glaciers and ice-fields. The topographic relief of the Queen Charlotte Mountains

[1]Transactions of the Fourteenth British Columbia Natural Resources Conference, 1962 (p. 28).

exceeds 3,000 feet and the highest peaks in the Vancouver Island Mountains reach elevations of more than 7,000 feet. Vancouver Island Mountains are separated from the Coast Mountains by the Georgia Depression, which is bordered by the Nanaimo Lowland on the west and the Georgia and Fraser Lowlands on the east. These Lowlands comprise the humid coastal belt.

The Eastern system includes the Rocky Mountains and the Rocky Mountain Foothills belt that form a continuous belt 100 miles wide stretching north from the International Boundary and limited on the west by the Rocky Mountain Trench. Many of their highest peaks are more than 11,000 feet in elevation and support alpine glaciers.

An intermediate unit, the Interior system, comprised of several major and minor mountain ranges and plateaux separates the eastern and western systems. It occupies a belt that averages more than 200 miles wide and extends southeast from Alaska to the southern boundary of British Columbia. The south part of the interior system includes the Columbia Mountains and the extensive Interior Plateau; the central and north parts contain several mountain ranges with plateaux and basins.

Climate and Water Balance

The Cordilleran hydrogeological region, because of its proximity to the Pacific Ocean and its high relief, has all types of climate from humid to polar. Mean annual temperatures decrease towards the north and changes with increasing latitude are shown in Table XVI.

Table XVI

Change in Temperature with Increasing Latitude

Station	Latitude	Mean Annual Temperature (°F)	Mean January Temperature (°F)	Mean August Temperature (°F)
Abbotsford	49° 1′	49°	33°	62°
Lytton	50°16′	51°	28°	71°
Dog Creek	51°38′	39°	13°	59°
Williams Lake	52°10′	43°	21°	63°
Ft. Nelson	58°50′	30°	−6°	58°

Air masses that originate in the Pacific and Arctic regulate the climate; they inherit the temperatures of their source areas but their direction and speed of movement are controlled by changes of atmospheric pressure and topography. Arctic air masses frequenting the area in winter months are slightly modified as they follow longitudinal valleys in their move southward. The Pacific air masses, common throughout much of the year, move eastward to encounter first the high mountains of the Western system that act as barriers to the movement of the air masses, which are forced to ascend and in so doing are cooled and loose their moisture as heavy precipitation along the coastal belt. Isohyetal lines on Figure 57 outline precipitation areas in

the Cordillera and in some places the Coast Mountains receive more than 100 inches annually. At the heads of long fiords, precipitation greatly exceeds this amount and the highest recorded on this continent is 251.3 inches at Henderson Lake at the head of a valley off Barkley Sound, west coast of Vancouver Island. Precipitation for the year 1963 at Ocean Falls totalled 171.6 inches and at Kitimat 120.4 inches.

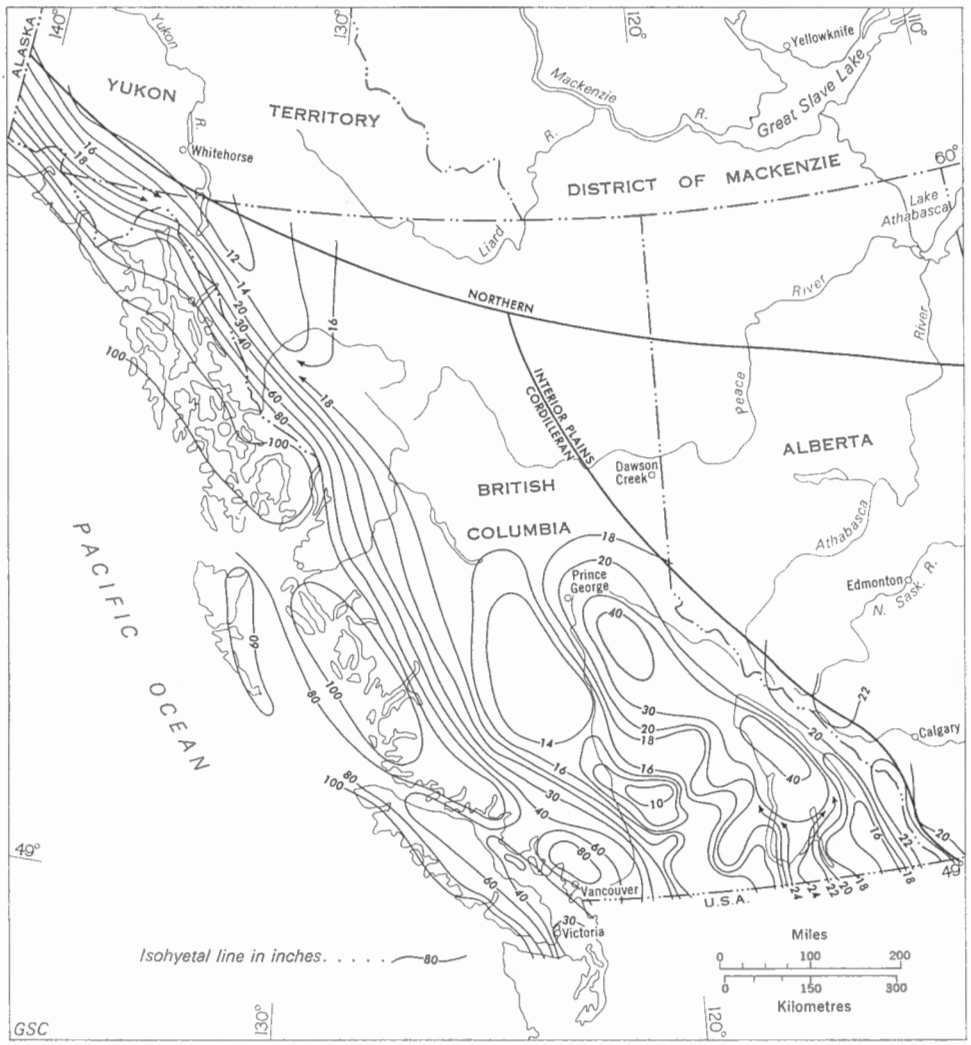


FIGURE 57. Mean annual total precipitation of the Cordilleran hydrogeological region.

After crossing the mountain barrier, drier air, in some instances heated, descends on the lee side. This produces an interior dry belt with precipitation rates varying from 12 to 16 inches annually; in places such as Kamloops in the Thompson Valley and Oliver in the Okanagan Valley, less than 10 inches are recorded annually.

In the south part of the Cordillera, the eastward moving air masses reach the Columbia and Rocky Mountains and again are forced to rise. This produces a second rain belt and annual average precipitation exceeds 40 inches in places. Upon crossing the Rocky Mountains the air mass again becomes drier and under certain conditions warmer. Thus, the eastward moving air becomes progressively drier inland with each windward slope having more precipitation than the lee side.

Precipitation is the basic source of water supply, and in the water balance losses are attributed to run-off, evaporation, evapotranspiration, and groundwater storage changes. Run-off, which accounts for a large percentage of the water losses, is recorded at about 475 stations mostly in the south part of the Cordillera. Table XVII shows precipitation–run-off relations for some basins in which records have been published. Run-off exceeds precipitation in some basins and this may be due to one or more of the following factors: (1) incorrect assessment of the drainage area, (2) substantial increase in run-off due to snow and glacier melt, (3) inadequate records of precipitation may not be representative of the entire basin.

Table XVII

Precipitation–Run-off Relations for some Basins in the Cordillera

River Basin	Drainage area (sq. mi.)	Mean Annual Precipitation (in.)	Mean Annual Run-off (in.)	Rock Types[1]
Koksilah	86	48	34	i, p
Zeballos	68	162	105	i
Browns River	33	60	51	i, p
Goat River	430	19	32	i, p
Keremeos	70	10	5	i, p
Nikomekl	38	57	45	p
Chilliwack	131	80	65	p, i
Skeena	15,000	19	33	i, g
Stikine	11,300	19	17	i, g

[1]p—pervious, i—impervious, m—mixed, g—glaciers.

During the summer months contributions to run-off may be negligible and losses by evaporation and evapotranspiration may exceed precipitation. Pan evaporation studies have been carried out at Agricultural Research stations for periods of several years and the values of these losses at some stations are compared in Table XVIII with precipitation for the same period.

Evaporation exceeds precipitation during the summer growing season at all stations excepting Agassiz, which is in the coastal humid belt. The losses during this season are from groundwater storage.

Run-off from mountains, evaporation, and evapotranspiration in the plateaux and valleys account for substantial losses in the water balance. With all these losses, the only source for groundwater is the heavy winter precipitation which in most places accumulates as snow. Although snowmelt is chiefly lost by direct run-off, part of it passes through a groundwater storage phase.

Table XVIII

Precipitation and Evaporation Loss during Summer Months

Station	May	June	July	August	September	October	Total
Saanich (13 yrs)[1]							
precipitation[2]	1.40	1.17	0.73	0.93	1.20	3.52	8.95
evaporation	3.72	4.06	5.01	4.04	2.66	1.26	19.49
Agassiz (14 yrs)							
precipitation	3.24	3.21	1.89	2.06	3.82	7.41	21.63
evaporation	3.00	2.89	3.95	3.42	2.16	1.60	15.42
Summerland (32 yrs)							
precipitation	1.07	1.35	0.96	0.80	0.77		4.95
evaporation	5.17	5.07	6.20	5.36	3.34		25.14
Prince George (15 yrs)							
precipitation	1.56	2.37	2.40	2.99	1.88		11.20
evaporation	2.85	3.06	2.99	2.59	1.57		13.06
Smithers (10 yrs)							
precipitation	1.45	1.62	2.07	1.12	1.67	2.02	7.93
evaporation	3.58	3.64	3.88	3.15	1.75		16.00

[1]Number of years measurements made.
[2]Precipitation and evaporation reported in inches.

Hydrogeology

Cordillera

The delineation of aquifers is essentially a description of rock types. The Cordillera includes rocks that were involved in folding, faulting, and mountain building as well as subsequent erosion and sedimentary deposition in broad basins. The geology is complex and the rocks are grouped in ages ranging from Precambrian to Recent. Aquifers or groundwater reservoirs include only those rocks that are permeable and hence provide pore spaces for the storage and movement of groundwater. Joints, fractures, and fault zones are common to all rocks without recognition of geological age and allow for the movement of water as is apparent at many bedrock springs.

In much of the Cordillera (Fig. 58, *in pocket*), the groundwater potential is unknown. The areas are mountainous, generally forested to elevations of more than 5,000 feet above which peaks are bare or capped with ice-fields. Surface water supplies are sufficient to meet the demands of the scattered lumber, fishing, and mining camps but no records are available indicating successful completion of wells in bedrock that yield large volumes of water.

In some mining camps, especially where limestones are present, groundwater percolating through solution channels has caused flooding and abandonment of mining operations. On the other hand, some mining camps located on major fault zones have experienced no problems of groundwater flooding (Table XIX). All the mines are damp and report some groundwater, but in most instances it is easily

Table XIX

Water Pumped from some British Columbia Mines

(From data compiled by Mineral Economics Section, Mineral Resources Division, Dept. of Mines and Technical Surveys)

Mine	Company	Location	Ore Mined	Depth (ft)	Major Mining Method	Water Pumped (gpd)	Remarks
Jersey	Canadian Exploration, Limited	Salmo	Lead–zinc	800	Open stoping	Low	Natural drainage
Bluebell	Cominco Ltd.	Riondel	Silver–lead–zinc	1,110	Open stoping	5,900,000	About half thermal H_2O, contains CO_2
H.B.	Cominco Ltd.	Salmo	Zinc–lead	1,000	Open stoping	Low	Natural drainage
Sullivan	Cominco Ltd.	Kimberley	Lead–zinc	2,000	Open stoping	Low	Part natural drainage
Highland-Bell	Mastodon-Highland Bell Mines Limited	Beaverdell	Silver	1,700	Open stoping	Low	Natural drainage
Reeves MacDonald	Reeves MacDonald Mines Limited	Remac	Zinc–lead	2,300	Open stoping	Low	Dry mine
Mineral King	Aetna Investment Corporation Limited	Toby Creek	Silver–lead–zinc	1,250	Open stoping	Low	Natural drainage
Bralorne	Bralorne-Pioneer Mines Limited	Bralorne	Gold	5,200	Cut-and-fill	Low	
Gold Quartz and Island Mountain	The Cariboo Gold Quartz Mining Company, Limited	Wells	Gold	2,000	Cut-and-fill	Low	
Giant Nickel	Giant Mascot Mines, Limited	Choate	Nickel–copper	1,200	Open stoping	Low	Natural drainage
Britannia	The Anaconda Company (Canada) Ltd.	Britannia Beach	Copper–zinc	4,000	Cut-and-fill	6,000,000	Estimated peak flow in May
Coast Copper	Coast Copper Company, Limited	Benson Lake	Copper	700	Shrinkage	Negligible	Natural drainage
Sunro	Cowichan Copper Co. Ltd.	River Jordan	Copper	550	Open stoping	Negligible	Natural drainage

controlled by pumping and in others it is utilized in sprinkling for dust control. Groundwater moves along bedding planes of flat-lying, gently dipping, or folded sandstones, conglomerates, and shales in sedimentary basins, and commonly sufficient water is obtained for domestic use from drilled wells. The operation of coal mines in many of these basins has been curtailed by groundwater inflow and upon cessation of mining the lower levels were quickly flooded.

Plateau lavas in the central part of the Interior system are potential aquifers probably recharged by large volumes of groundwater from snowmelt. The discharge of groundwater by springs at the base of the lavas maintains the flow of many streams during the seasons of little or no run-off.

The unconsolidated surficial deposits collectively defined as drift provide the best natural facilities for the storage of water. These deposits are of glacial origin and were deposited during Pleistocene time when ice-sheets covered most of the Cordillera. The distribution of glaciers and ice-fields that exist today is shown on Figure 58. During Pleistocene time, ice-fields accumulated on these same peaks and developed into small valley glaciers that with continued growth filled the major valleys and advanced into the interior lowlands. Eventually enough ice had accumulated to form an ice-sheet that covered most of the Cordillera. Ice-sheets of continental proportions accumulated at least three times during the Pleistocene, and during the periods between glaciations climatic conditions much the same as today prevailed. During the last glaciation, some of the higher peaks, especially those above an elevation of 8,000 feet, remained above the ice-sheet. At its maximum, the Cordilleran ice-sheet was controlled by climatic factors rather than topography and it expanded eastward to meet at the Rocky Mountain Foothills belt the Laurentide ice-sheet that occupied central and eastern Canada. The ice advanced to a point about 140 miles south of the International Boundary and on the west it discharged into the Pacific Ocean.

Eventually climatic conditions changed and the Cordilleran ice-sheet disappeared through downmelting and decay rather than by retreat. This was not a continuous process but at times ice-tongues were nourished and valley glaciation resulted locally. Apart from sculpturing and moulding the mountains and their peaks, the advancing ice disrupted the drainage systems and eroded and overdeepened the major valleys. The most widespread of all the materials left after disappearance of the glaciers was ground moraine or till that mantles the lower slopes, valleys, and plains. In places, it rests directly on bedrock and elsewhere overlies unconsolidated sediments deposited during interglacial periods. Ground moraine or till consists of compact, unsorted mixtures of sand, silt, clay, and stones and does not constitute a favourable groundwater reservoir. Large volumes of meltwaters released as the ice melted were loaded with glacial debris that filled valleys and in places spread as outwash gravel and sand over considerable areas. It is commonly well sorted and constitutes the most important reservoir both for recharge and storage of groundwater. Blocks of stagnant masses of ice, lying in valleys, ponded water to form ice marginal lakes into which silts and clays were deposited in places to depths of 500 feet or more. On the coastal lowlands, eustatic changes of sea-level provided marine environments for the deposition of thick sections of clay, and in places, spits and beach ridges were deposited to mark former higher sea-levels.

Unconsolidated deposits of glacial, glacio-fluvial, lacustrine, alluvial, and wind-blown materials mantle the floors and lower slopes of the intermontane valleys. The present rivers have entrenched themselves into these deposits in places to depths of 200 feet or more and the overdeepened portions of some valleys contain lakes. The valleys vary from 1 mile to 10 miles in width and provide soils for agriculture; hence they have given rise to population centres and transportation routes.

Permeable outwash gravel and sand that is perhaps more than 300 feet thick fills the Rocky Mountain Trench. Drill-holes have penetrated as much as 250 feet of these permeable unconsolidated deposits without encountering bedrock. Not all valleys are filled with permeable sediments and more than 300 feet of till has been reported as filling the Clearwater and North Thompson valleys. At Enderby in the Okanagan Valley, test wells have penetrated more than 1,300 feet of unconsolidated sediments. Not all the sediments filling valleys are related to the last glaciation but some may represent fluvial deposits of interglacial times when climatic conditions were much the same as at present, and these may in turn be underlain by drift deposited by an earlier glaciation.

Humid Coastal Lowland

The humid coastal lowland includes the Fraser Lowland and the Georgia and Nanaimo Lowlands that border the Gulf of Georgia. They are the most densely populated areas of the Cordillera and the demand for water there is greater than elsewhere. Although surface water meets the requirements of the metropolitan areas, groundwater augments supplies for areas beyond main pipelines and it is also in demand for irrigation and industrial use.

The Fraser Lowland, or as it is commonly called 'the lower mainland of south-western British Columbia', extends west from the village of Hope to the Gulf of Georgia. It lies between the Coast Mountains on the north, the Cascade Mountains on the south and east, and its south boundary, west of Vedder Mountain, is the International Boundary. The distance from Hope to downtown Vancouver is 100 miles. The Fraser Lowland consists of wide relatively flat-topped uplands and wide flat-bottomed valleys with a through east-west post-glacial valley occupied by Fraser River. Tributaries of the Fraser drainage system carry run-off from the bordering Coast and Cascade Mountains. Within the lowland, Nicomekl, Salmon, Sumas, and Serpentine Rivers, Campbell, Bertrand, Fishtrap, Willband, Clayburn, and Nathan Creeks, as well as many smaller creeks, carry run-off during periods of excess rainfall and at other times their flow is maintained by groundwater discharge that is estimated to exceed 30 million gpd.

The unconsolidated surface deposits are commonly more than 600 feet thick and evidence from deep test holes drilled for oil and gas indicates thicknesses of as much as 1,800 feet at Tsawwassen Indian Reserve, Boundary Bay; 1,047 feet at Pitt Meadows; 815 feet at Surrey Center; more than 1,010 feet at Langley; 1,105 feet at Abbotsford airport. The Pleistocene geology is complex but at least three major glaciations that reached ice-sheet proportions took place in the area, and a fourth, primarily a valley glaciation, was responsible for the distribution of outwash gravel and sand that constitutes the principal recharge areas in the eastern part of the lowland. The principal

waterbearing reservoirs include a) outwash gravel and sand; b) fluvial sand and gravel between two stony clays; c) fluvial sand and gravel confined beneath ground moraine or till; and d) interconnected lenses of coarser material included in the silt, clay, and fine sand that fills, in places, the Nicomekl–Serpentine and Langley Valleys to depths of more than 900 feet. Flowing artesian groundwater conditions exist in those confined aquifers in the valleys, and flows of as much as 200 gpm are recorded but commonly less than 15 gpm are encountered by small diameter bores.

The source of groundwater is precipitation, the distribution of which is not uniform but increases northward and eastward, so that Vancouver airport at Sea Island receives an average of 44 inches annually whereas Agassiz at the east end of the valley gets an average of 64.39 inches. Another characteristic of the rainfall pattern is that less than 7 per cent of the annual total falls in July and August. Recharge is effective during the winter rainy season and the response of water levels to precipitation is noted on Figure 59 where static levels in two observation wells are compared to local precipitation. Nearly all communities and farms south of Fraser River rely on groundwater for domestic use and many crops are irrigated by it. It is estimated that groundwater consumption in that part of the Fraser Lowland exceeds 20 million

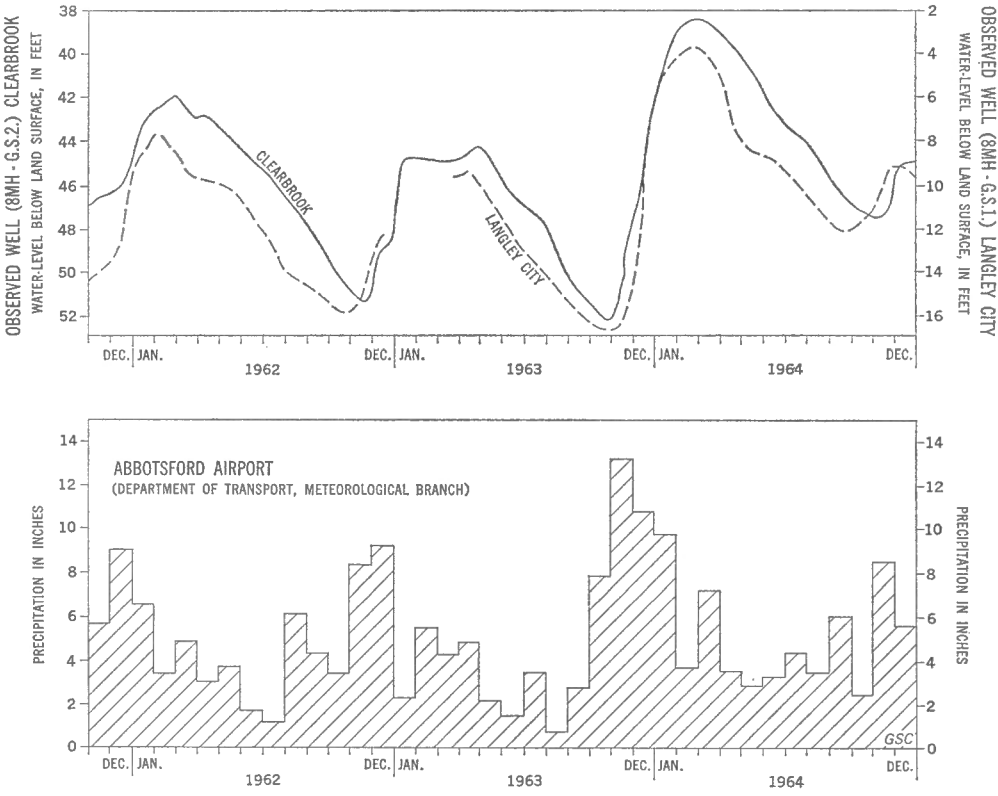


FIGURE 59. Observation well hydrographs, Fraser Valley, British Columbia.

gpd. A test well, drilled by the Geological Survey of Canada at Aldergrove in the central part of the lowland, penetrated 759 feet of unconsolidated deposits. Two major aquifers were encountered, the lower of which is confined beneath till and is under observation to determine groundwater behaviour in a confined aquifer.

The unconsolidated deposits of the Georgia Lowland consist of marine clays that overlie till or lie directly on bedrock. In places, a thin mantle of gravel or sand that accumulated during higher sea-levels constitutes local aquifers, and supplies sufficient for domestic use are available. In general, groundwater is not available in sufficient quantities and it is necessary to store surface run-off.

Thick unconsolidated deposits constitute the groundwater reservoirs of the Nanaimo Lowland. Till, the most widespread surficial deposit, mantles much of the lowland and in places is separated from the underlying bedrock by fluvial, marine, and glacial materials as much as several hundred feet thick. Overlying the till and extending to the ground surface, the principal groundwater reservoirs include a) ice-contact gravel and sand of varied form; b) deltaic gravel and sand accumulated where rivers reached former higher sea-levels; and c) modern river bottom, deltaic and alluvial fan deposits. These deposits, recognized by their physiographical form, are rarely more than a mile or two across but many of them are several tens of feet or even 100 feet thick. The water-bearing materials beneath the till are principally fluvial sands and gravels that were formed during non-glacial intervals prior to the last glaciation. They are more extensive and uniform than the aquifers above the till, but the thickness and position of the water-saturated material within them fluctuate greatly from place to place with variation in the position of the irregular impervious layers beneath and above them. In some places, two or more distinct water-bearing series of gravel and sand are separated by impervious silt, clay, or till but generally only the uppermost till is present. This subtill succession of horizontally stratified sand and gravel constitutes the largest groundwater reservoir of the lowlands and large capacity wells tapping it supply water systems at Comox, Bowser, Qualicum, Lantzville, Cowichan Bay, and Sidney.

Hydrochemistry

The results of chemical analyses of groundwaters have been published in reports and bulletins prepared by officers of the Geological Survey. Also the Industrial Minerals Division, Mines Branch, has reported on civic water supplies as well as surface waters in drainage basins covering the Cordillera. All the reports are listed herein with the selected references.

Chemical analyses presented in Table XX include the constituents as analyzed in parts per million for samples representing unconfined aquifers in outwash areas, confined aquifers beneath tills, and two examples of water from bedrock. One of the samples representing bedrock waters was collected from a spring on Saltspring Island and the other sample was taken from the top of an abandoned coal mine shaft. The quality of the water in the mine shaft may not be represented there because of probable inflow near the surface of groundwater from permeable surficial deposits.

Table XX

Chemical Analyses, Cordilleran Hydrogeological Region

Sampling Point	Location	pH	Conductance	Constituents Analyzed (parts per million)															
				Non-carbonate hardness	Total hardness	Alkalinity as $CaCO_3$	Calcium	Magnesium	Sodium	Potassium	Total iron	Manganese	Bicarbonate	Sulphate	Chloride	Fluoride	Nitrate	Silica	Sum of constituents
UNCONFINED AQUIFERS																			
Well, 50 ft deep	Long. 122°41′ Lat. 49°04′21″	7.5		2.6	37.2	34.6	9.8	3.1	2.4	0.7			42.2	4.9	2.1	0.0	1.2	17.5	62.6
Well, <50 ft deep	Long. 128°43′ Lat. 60°4′	7.7	346	5.5	168.0		49.5	10.8	5.8	1.2	0.27	0.04	198	7.2	10.2	0.06	4.7	14.0	201
CONFINED AQUIFERS																			
Well, 358′ deep	Long. 122°28′5″ Lat. 49°04′17″	8.5	1,293	0.0	29.2	557	2.4	5.6	302	6.7	0.37	0.06	679	23.6	74.6	1.4	0.0	24	775
Well, 113′	Long. 122°44′48″ Lat. 49°09′35″	8.1	267	0.0	53.2	122	13.9	4.5	36.0	4.5	0.01		148	3.2	9.7	0.0	0.0	41	186
Springs	Long. 124°20′ Lat. 49°18′43″	7.6	163	0.0	72.9	79.4	16.6	7.7	5.0	0.5	0.08		96.8	1.6	3.0	0.0	2.0	25	109
BEDROCK																			
Mineshaft	Long. 123°58′53″ Lat. 49°12′20″	7.8	496	0.0	67.8	242	18.1	5.5	90	1.3	2.4	0.11	295	7.7	11.2	0.16	0.8	18	298
Saltspring	Long. 123°32′24″ Lat. 48°54′31″	7.0	70,690	1,458	2,154	696	510	214	25,240	580	0.66	0.20	849	17,520	27,440	1.5	0.3	2.5	71,926

In general the groundwaters are low in mineralization and hardness. As the groundwaters increase in hardness there is a tendency for the iron and, at times, the manganese to increase. Groundwaters, unsatisfactory for most uses, do appear in certain geological environments, especially where leaching produces a concentration of salts; where oxidation increases the iron content; and in the coastal areas where bedrock wells may become contaminated with saltwater encroachment due to overpumping.

Thermal Springs

Thermal springs as defined by Elworthy (1925) are those springs with temperatures ranging from 70°F to 90°F, and hot springs have temperatures of more than 90°F. All springs with temperatures greater than the normal groundwater temperatures of the area are here classified as thermal springs. Normal groundwater temperatures of 52°F are recorded at Aldergrove (lat. 49°04′) and of 48°F at Vanderhoof (lat. 54°01′). Thermal springs, Figure 58, are found in fault zones and are aligned coincident to or parallel with the general northwest structurally controlled topographic features. Thermal springs on the flanks of the Coast Mountains are commonly near centres of Tertiary and Recent volcanism. The source of the thermal springs is groundwater that percolates downward through fracture zones, is heated, and then discharged at the surface. Commonly at all the spring localities more than one spring will be found but all are believed to emanate from the same fissure or fracture zone.

Most of the thermal springs in the south part of the Cordillera were sampled by Elworthy in 1923 and he found that calcium sulphate and calcium bicarbonate were the chief constituents and hydrogen sulphide was released when the waters emerged into the open air. Sampling of a spring at Banff during an eight-year period indicated little variation in the chemical composition and little change in temperature. In the Ainsworth district, at the Bluebell mine, on July 4, 1962, a heavy flow of water and carbon dioxide from a drill-hole forced abandonment of the lowest level of the mine.

Table XXI

Thermal Springs in the Cordilleran Hydrogeological Region

Locality	Reported Temp. (°F)	Chief Constituents	Remarks
Banff	100	Ca, Mg, SO_4	5 springs
Radium	108–118	Ca, SO_4	3 springs
Fairmont		Ca, Mg	several springs
Halcyon	123	Ca, Na, SO_4	3 springs
Ainsworth	101	Ca, Na, HCO_3	
Jasper	120	Ca, Mg, HCO_3	several springs
Harrison	140–145	Ca, Na, SO_4	2 springs
Lillooet		Mg, SO_4	3 springs
Vancouver Island	125	Cl	Sharp Point Hot Spring
Coast area	100	Na, HCO_3	several springs
Lakelse	187		contains lithium
Edziza	150–195		several, deposits tufa
Liard River	110–121	Ca, SO_4	3 springs

The water at a temperature of 105°F flowed at a rate of 400 gpm with a pressure of 325 pounds per square inch. In the Edziza Peak–Spectrum Range area and Mess Creek valley, Souther[1] reports that some of the thermal springs are actively depositing calcareous tufa and destroying the present vegetation. None of these hot springs is boiling but some appear to have temperatures ranging from 150°F to 195°F.

Many springs have been recorded as mineral springs that discharge at normal temperatures. The known thermal springs of the Canadian Cordillera are listed in Table XXI.

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Chapter VIII

NORTHERN HYDROGEOLOGICAL REGION

by E. B. Owen

Introduction

The northern region includes, in general, all of Canada north of the southern limit of discontinuous permafrost (Fig. 60, *in pocket*). Throughout this vast region data are scarce, precipitation is low, and permafrost is generally present. Thus, for the present, it would serve no purpose to subdivide it despite the great variety of topography and geology, and it is only possible to make broad generalizations as to the availability of groundwater. An estimate of groundwater supplies is hindered by the absence of wells and other subsurface data throughout most of the area. Some water well drilling data are available for several small communities along the Alaska Highway and in the Mackenzie River drainage basin, but information is sparse for the Quebec–Labrador region east of Hudson Bay. The data available on climate, run-off and geology are, however, sufficient to indicate that groundwater is probably present throughout the areas of discontinuous permafrost and that the rocks are saturated in the same manner as in more humid, temperate regions.

Definitions peculiar to these conditions follow.

Permafrost or permanently frozen ground is defined by Muller (1947) as "a thickness of soil or other superficial deposit, or even a bedrock, at a variable depth beneath the surface of the earth in which a temperature below freezing has existed continuously for a long time (from two to tens of thousands of years)."

Dry permafrost is ground whose temperature is permanently below freezing but contains no ice as a cementing substance. Dry permafrost is usually found in sandy or other coarse-grained, clastic material that drains easily.

The *active layer* is the thin surface layer of the ground that freezes in winter and thaws in summer. An equivalent term is *seasonally frozen ground* (Muller, 1947).

A *talik* is a layer of unfrozen ground within the permafrost. The term also applies to a layer of unfrozen ground between the active layer and the permafrost and to the unfrozen ground beneath the permafrost (Muller, 1947).

The accompanying map (Fig. 60) indicates the approximate southern limits of continuous and discontinuous permafrost in Canada (Brown, 1963). In the continuous zone, permafrost occurs everywhere beneath ground surface and is generally

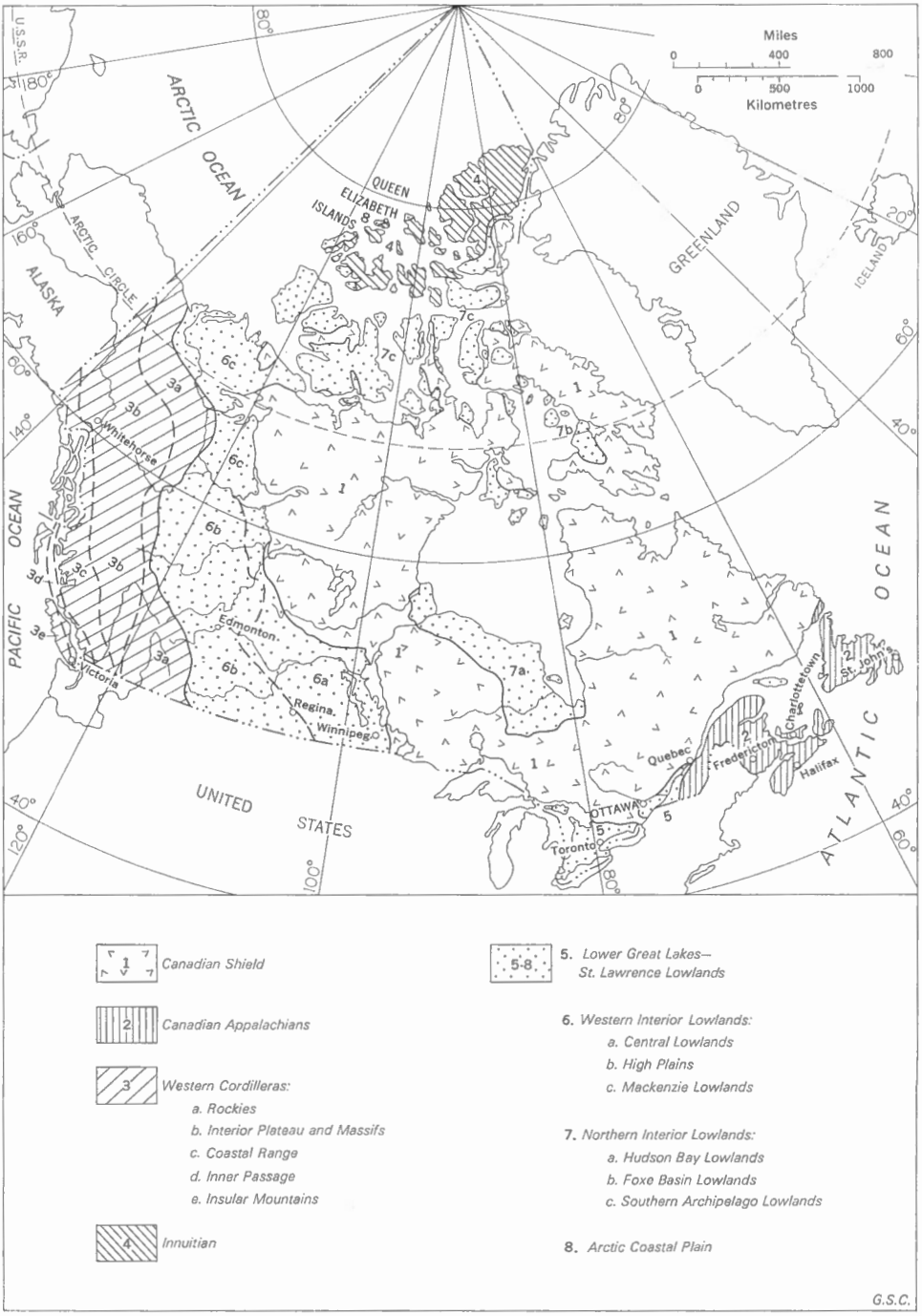


FIGURE 61. Physiographic regions of Canada.

hundreds of feet thick. Southward the continuous zone grades into the discontinuous zone, where permafrost exists in combination with areas of unfrozen material. The discontinuous zone is one of broad transition between continuous permafrost and ground having no permafrost. In its southern part, permafrost occurs only in scattered patches and is only a few feet thick. Permafrost also exists at high altitudes in southern Labrador-Ungava and in the mountains of western Canada a great distance south of the limit of discontinuous permafrost shown in the figure.

Physiography

The vast area in which permafrost exists in Canada extends across the northern parts of three physiographic regions: the Canadian Shield, the Interior Plains, and the Cordillera (Fig. 61).

The largest of these is the Canadian Shield, which occupies nearly half of Canada including parts of the Arctic Archipelago. As a whole, the Shield consists of a rocky plateau with old worn-down mountains above and enclosed plains below its general surface. Its relief ranges from 10,000-foot-high mountains on Baffin Island to a few feet in the part surrounding Hudson Bay. In most places the Shield has a relief of less than 200 feet. The whole of the Shield has been glaciated. The ice-deepened pre-existing valleys wore down ridges and spread quantities of debris. Large temporary lakes were created that left behind deposits of lacustrine clay and beach sand and gravel. Thick deposits of sand and gravel were formed in many of the valleys by melt-waters flowing from the wasting ice-sheets.

The Mackenzie Lowland, which forms the north part of the Interior Plains region, is some 1,100 miles long and up to 300 miles wide. For the most part it consists of a series of plateaux ranging from an elevation of 500 feet in the east to 4,000 in the west. Long, deeply entrenched tributaries to Mackenzie River, such as the Athabasca, Peace, Arctic Red and Peel Rivers, enter it from the west. Extensive glacio-lacustrine silt and clay occur in the eastern part of the Mackenzie Lowland around Lake Athabasca and Great Slave and Great Bear Lakes. These were deposited by large lakes that existed in these areas during glacial times. Interbedded with these materials are deltaic deposits of sand formed by rivers emptying into the glacial lakes. Bedrock formations of the Mackenzie Lowland consist mostly of sedimentary rocks of Palaeozoic or later age.

As it crosses the Cordilleran region, the southern limit of permafrost closely follows the 60th parallel, which is the boundary separating Yukon Territory and the province of British Columbia. In this region the area includes parts of three physiographic provinces, namely, the Coast Range province on the southwest, the Yukon Plateau province in the south-central part, and the Mackenzie Mountain province to the north and east of the plateau. Compared with the nearly flat topography of the Mackenzie Lowland, which borders it on the east, the area is mountainous and contains a wide variety of rocks with exceedingly complex structures.

The Coast Mountains constitute a wide upwarp cut into blocks by deep, U-shaped valleys, many of which traverse the mountains completely. The St. Elias

Range constitutes the greater part of the Coast Mountains in the area where permafrost exists. Here active glaciers have cut deep troughs and sharp ridges. Mount Logan, elevation 19,850, the highest peak in Canada, is in this range.

The Yukon Plateau consists of a gently rolling upland surface dissected by valleys and larger areas cut well below it. Its elevation seldom exceeds 5,000 feet. The plateau is interrupted by individual mountains and by ranges of various sizes, the largest of which are the Pelly Mountains. The greatest elevation in this range is about 8,000 feet.

The Mackenzie Mountains area comprises mountains, plateaux, and plains, containing within it such diverse features as Liard Plateau, the Franklin Mountains, Mackenzie Plain, the Mackenzie Mountains and Peel Plateau. The rocks are almost wholly sedimentary. In general, the area can be described as a basin floored by broad plains and low hills and rimmed by mountains. The rim of the basin is formed in part by the Mackenzie, Wernecke, and Ogilvie Mountains on the south and southwest, the Richardson Mountains on the east, and the British Mountains on the north. The main feature of the basin is Porcupine Plateau, a gently undulating, lake-spotted plain drained to the west by Porcupine River. Another large river is the Peel, which flows east past the south end of Richardson Mountains and north to Mackenzie River. There has been a general lack of glaciation in the area. Ice accumulated in the Wernecke Mountains moved north across the south part of Porcupine Plateau; other ice moving northwestward from the Mackenzie Valley was, in part, diverted northward along the east slope of the Richardson Mountains.

Climate

The area contains three climatic regions. One comprising the southern half has been classified as *subarctic*; i.e. it has cool short summers with only one month to three months with mean temperatures above 32°F. The second region is *tundra*, where the mean temperature of the warmest month is below 50°F but above 32°F. Here, the climate is characterized by extremely short and cool summers and very long, cold winters. The dividing line separating the subarctic and the tundra regions, i.e., the 50°F mean temperature isotherm for July, closely approximates the northern limit of tree growth. It also coincides generally with the dividing line between the continuous and the discontinuous areas of permafrost. The third region is *ice-cap*, where the mean temperature for all months is below 32°F. This region, which is relatively small, occupies mountain peaks in the Cordillera and places in the north and west parts of the Arctic Archipelago.

Monthly and annual averages of daily mean temperatures and total precipitation in inches for twelve selected weather stations are listed in Tables XXII and XXIII.

Hydrogeology of Permafrost Areas

The occurrence of groundwater in permafrost areas differs somewhat from its occurrence in warmer climates because of the presence of frozen, impermeable ground underlying or interbedded with unfrozen deposits with varying degrees of permeability. In the area of discontinuous permafrost, groundwater may be obtained in unfrozen

Table XXII

Monthly and Annual Averages of Daily Mean Temperatures, Northern Hydrogeological Region (°F)

(From Meteorological Branch Records, Department of Transport)

Location Month:	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
Aklavik, N.W.T.	−18	−17	−8	9	31	49	56	50	38	20	−3	−16	16
Coppermine, N.W.T.	−19.4	−20.5	−13.8	1.0	21.8	38.5	49.0	46.6	36.3	19.3	−3.8	−15.2	11.7
Baker Lake, N.W.T.	−31	−27	−13	−2	23	36	49	50	36	17	−3	−18	10
Fort Chimo, Que.	−13	−12	3	16	32	45	53	50	42	31	17	0	22
Dawson, Y.T.	−16.0	−11.4	−6.0	28.4	46.6	57.0	59.8	54.6	43.1	27.4	0.20	−1×2.8	23.7
Whitehorse, Y.T.	5.2	6.8	21.3	31.8	45.5	54.6	56.2	53.6	46.0	34.4	14.6	3.2	31.1
Watson Lake, Y.T.	−7.6	−1.2	16.0	31.2	46.5	56.4	58.7	55.0	46.2	34.0	8.3	−7.8	28.0
Yellowknife, N.W.T.	−14.7	−14.2	1.4	17.3	38.9	53.3	60.9	56.7	44.9	31.0	7.2	−12.9	22.5
Fort Smith, N.W.T.	−13.4	−8.7	5.4	26.2	45.2	55.4	61.2	56.8	44.2	31.7	9.8	−7.0	25.6
Churchill, Man.	−18	−17	−4	12	28	42	53	53	42	30	11	−7	19
Fort Mackenzie, Que.	−13	−11	1	16	33	46	54	53	44	33	18	−2	23
Knob Lake, Que.	−13	−8	8	24	35	48	55	52	42	30	15	2	24

Table XXIII

Monthly and Annual Averages of Total Precipitation in Inches, Northern Hydrogeological Region

(*From Meteorological Branch Records, Department of Transport*)

Location Month:	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
Aklavik, N.W.T.	0.55	0.49	0.38	0.50	0.49	0.80	1.39	1.42	0.91	0.86	0.76	0.42	8.97
Coppermine, N.W.T.	0.60	0.40	0.71	0.67	0.55	0.91	1.52	1.69	1.25	1.23	0.78	0.56	10.87
Baker Lake, N.W.T.	0.22	0.20	0.31	0.27	0.21	0.70	0.86	1.09	0.83	0.50	0.34	0.29	5.82
Fort Chimo, Que.	0.82	0.56	0.80	0.84	1.19	1.27	2.29	2.52	1.94	1.53	1.47	1.14	16.37
Dawson, Y.T.	0.88	0.62	0.62	0.37	1.06	1.24	1.72	1.81	1.31	1.07	1.04	0.99	12.73
Whitehorse, Y.T.	0.64	0.47	0.60	0.41	0.57	1.00	1.63	1.53	1.34	0.71	0.98	0.75	10.63
Watson Lake, Y.T.	1.30	1.03	0.68	0.79	1.03	2.00	1.87	1.91	1.64	1.26	1.63	1.61	16.75
Yellowknife, N.W.T.	0.54	0.46	0.39	0.36	0.59	0.73	1.15	1.02	0.92	0.99	0.69	0.61	8.45
Fort Smith, N.W.T.	0.53	0.66	0.67	0.52	0.97	1.41	1.99	1.65	1.54	0.96	0.86	0.87	12.63
Churchill, Man.	0.39	0.34	0.66	0.67	0.61	1.43	2.51	2.41	1.94	1.53	0.78	0.60	13.87
Fort Mackenzie, Que.	1.12	0.88	1.07	1.05	1.64	1.55	3.46	3.15	2.28	1.88	1.36	1.23	20.67
Knob Lake, Que.	1.39	1.67	1.98	1.32	0.90	3.33	3.76	2.86	2.85	3.21	2.95	1.32	27.54

zones. Farther north, however, in the area of continuous permafrost, owing to the absence of unfrozen ground, the development of groundwater supplies is frequently impractical if not impossible.

The main effect of permafrost on the hydrology of an area is that it restricts the movement of groundwater between drainage basins. In regions of low relief, water from surface run-off and the thawed water from the active zone are retained in surface depressions in the form of lakes and swamps. Conversely, owing to the absence of infiltration, run-off is rapid in areas of pronounced topographic relief.

Types of Occurrences of Groundwater

Groundwater may occur as (1) suprapermafrost water, (2) intrapermafrost water occurring in thawed zones between masses of frozen material (Fig. 62), or (3) subpermafrost water.

1. *Suprapermafrost water.* Nearly everywhere in permafrost regions the upper surface of the ground is thawed to some extent during the summer months. This is especially true in areas where the overlying layer of moss and decayed organic material is thin or non-existent. The thawing, if it extends deep enough, will create an appreciable reservoir of groundwater perched upon the underlying frozen, impermeable material. This is known as suprapermafrost water and is available for shallow wells during the summer months. Included with the suprapermafrost water is the groundwater that occurs beneath lakes and rivers. This water is more widely distributed and occurs in greater quantities than other groundwater existing in permafrost regions. Johnston and Brown (1963) described the existence of a thaw basin beneath a small lake in the Mackenzie River Delta near the town of Inuvik, Northwest Territories. Their drilling investigations showed the permafrost table sloped steeply downward from the edge of the lake, and the frozen ground present around the perimeter of the lake could be in the form of a wedge with its base curving back and downward toward the shoreline. The sediments beneath the centre of the lake were unfrozen to a depth of 230 feet.

Investigations by the United States Geological Survey of the thermal regime of a small lake near Barrow, Alaska, indicated that the permafrost table existed at a depth of 190 feet beneath the deeper part of the lake (Brewer, 1958). Ponomarev and Tolstikhin (1959) state that "suprapermafrost water is widely distributed beneath river beds and lake bottoms".

In some Canadian communities in the northern part of the area of discontinuous permafrost, wells have been dug into the active zone. These wells, however, are usually temporary, for during the winter the water saturating the ground above the permafrost table may become completely frozen. The exploitation of these wells could be prolonged by placing them beneath heated dwellings, where they would be partly protected. In some communities the water in the seasonally thawed layer can easily become polluted from nearby waste-disposal facilities and is not suitable for domestic use. The quantity available depends upon the permeability of the thawed material: the greater the permeability the deeper it thaws and the greater the abundance of water.

In general, the water in the numerous small springs and seeps that occur in

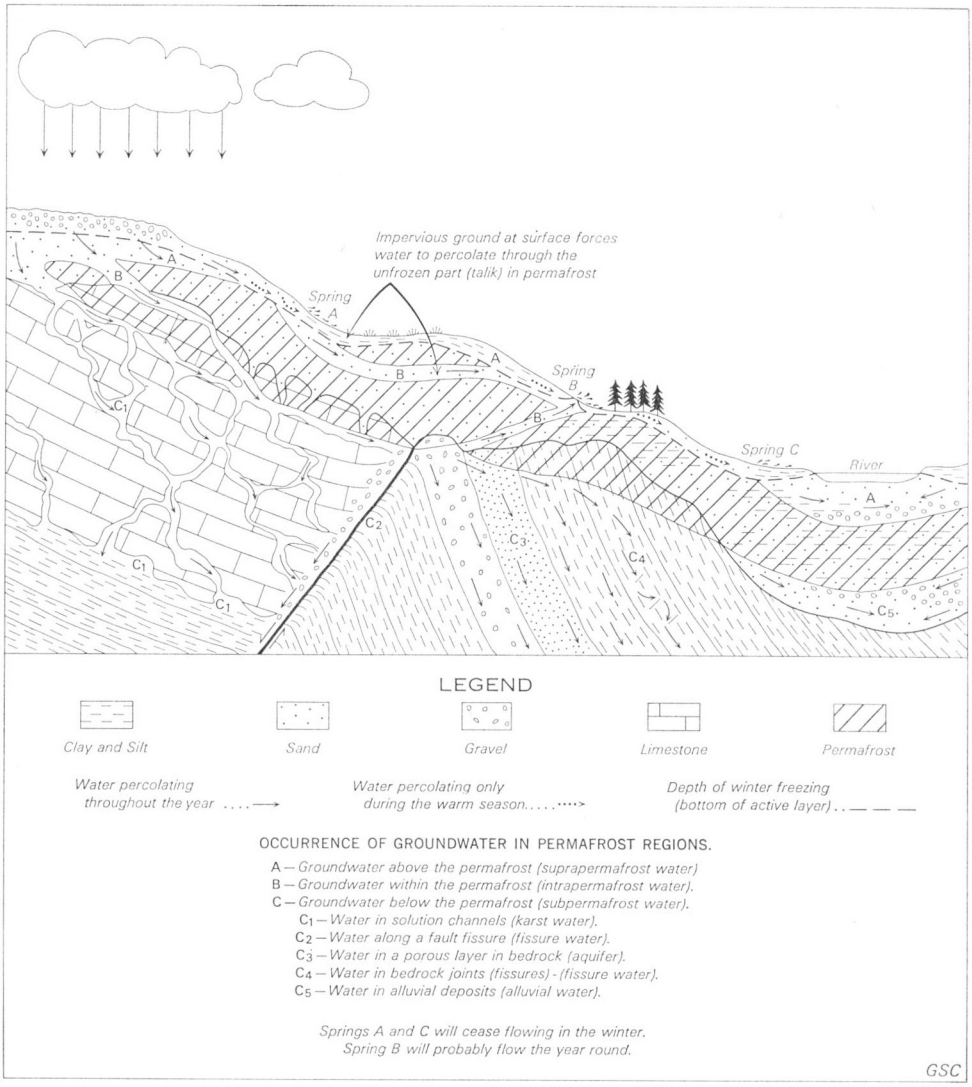


FIGURE 62. Occurrence of groundwater in permafrost areas.

summer along the edges of many of the streams and rivers in the permafrost regions would be classified as suprapermafrost water. The springs and seeps usually occur along the lower parts of south- and west-facing slopes and are intermittent, the greatest flow occurring in the afternoon and evening with little if any flow in the early morning. The frozen material from which the water is flowing usually consists of sand, gravel, or talus, the largest flows coming from the coarser-grained material. The mineralization of this water is usually low, although the organic content is sometimes high. The temperature seldom exceeds 40°F.

2. *Intrapermafrost water.* This exists within the thawed zones of permanently frozen ground. It commonly occurs in alluvium near rivers, in abandoned river channels, and in glacio-fluvial material covering the floor of wide river valleys such as Pelly River valley in Yukon Territory. In many places groundwater in the intrapermafrost zone is connected hydrologically with water in the supra- and subpermafrost zones. This is especially true in the central and southern parts of the area of discontinuous permafrost. In the more northerly areas the taliks, which are the aquifers carrying the intrapermafrost water, are less common and there is consequently less of this type of water available.

Wells drilled in such widely separated communities as Mile 1095, Alaska Highway, Mayo, Yukon Territory, and Hay River and Fort Simpson, in the Northwest Territories, all situated in the discontinuous permafrost area, obtain quantities of intrapermafrost water from coarse-grained materials between frozen materials. In some instances, especially where the wells are close to a stream, part of the water may be suprapermafrost.

Springs issuing from gravel underlying till frequently occur along the toes of bluffs along such northern streams as the Snake, Bonnet Plume, and Arctic Red Rivers, in Yukon Territory. This may be intrapermafrost or subpermafrost water. Little is known regarding the occurrence of permafrost in these areas except that it usually exists within a few feet of ground surface on the terraces above the bluffs. Much of the water in such rivers as the North Klondike, which has a year-round flow sufficient to provide a small hydro-electric power plant near Dawson, Yukon Territory, with a continuous water supply, may originate in intrapermafrost zones.

The mineralization of intrapermafrost water depends upon the conditions of formation of the taliks in which it occurs. The waters that rise to the surface from taliks that are connected to the active zone usually have a low mineral content, whereas a talik that is connected downward, possibly by some bedrock structure, to a deep-seated aquifer may contain highly mineralized water. This may account for the wide variation in the chemical analyses of groundwater that are included in Table XXVI.

3. *Subpermafrost water.* In a great many areas of Alaska subpermafrost water in the thawed zone beneath permanently frozen ground is a source of groundwater supply (Cederstrom, Johnston, and Subitzky, 1953). Many wells have penetrated 40 to 200 feet of permafrost in the Fairbanks area and have developed excellent permanent supplies of water beneath the frozen zone.

There is little information available concerning the occurrence of subpermafrost water in Canada. In the discontinuous permafrost areas the most likely places would be beneath the floors of valleys filled with thick deposits of fine-grained alluvium overlying coarse-grained material. If the valley fill is not frozen to bedrock subpermafrost, water should be available. As the impermeable frozen ground would act as a confining bed to water moving down from the higher ground along the sides of the valley, it is possible that artesian conditions exist beneath the valley. Examples are the flowing wells that occur along the Alaska Highway between Haines Junction and the Alaska boundary. Here the highway lies in Shakwak Trench, a great valley up to

10 miles in width, which follows along the northeast side of the St. Elias Mountains for a distance of about 200 miles. The floor of the valley is covered with extensive deposits of coarse-grained, glacio-fluvial sand and gravel. The water in the flowing wells is reported to be coming from beneath the permafrost (Brandon, 1965). Another example is an artesian well in Eldorado Creek valley near Dawson, Yukon Territory, which has been described by Tyrrell (1903). A 221-foot shaft was sunk to bedrock through frozen muck, sand and gravel. Artesian water was observed in the bedrock flowing at an approximate rate of 1,000 gpm. The aquifer in bedrock was assumed to be a fault encountered near the bottom of the shaft. The possibility of obtaining subpermafrost water in the areas of continuous permafrost is poor. Frequently the permafrost extends down to impervious bedrock so that no water is available. The mineral content of subpermafrost water is frequently high. The possibility of pollution is low.

Table XXIV indicates the volume of water pumped from nine Canadian underground mines situated in areas of discontinuous permafrost. As some of the water was introduced into the mine during mining operations, the figures should not be taken as representing the actual quantities of groundwater encountered. In some instances large volumes of water were encountered in the upper levels, which were subsequently sealed off. Most of the groundwater occurred along steeply dipping faults that apparently extended upward to bedrock surface. The quantity of water varies with the permeability of the overlying superficial deposits. Smaller amounts exist in mines where clayey material overlies bedrock than in others where the overburden consists of coarser-grained materials. The presence of permafrost probably has some effect upon the quantity of groundwater encountered in a mine. The volume of water pumped at United Keno Mines at Elsa, Yukon Territory, where the upper 300 feet of rock is frozen, is considerably less than that pumped from other mines where the permafrost is more discontinuous. Bateman (1949) states that at Giant Yellowknife Mines, in a mine considered to be relatively dry, permafrost existed to a depth of 280 feet.

In 1947 five wells were drilled at Fort Churchill, Manitoba, by the United States Army Corps of Engineers in search of a supply of potable groundwater for the army camp (Crumlish, 1948). In one well (No. 5) a small quantity of brackish water was obtained from fine-grained sand underlying the permafrost (Fig. 63). The water is believed to have been subpermafrost water contaminated by marine water from Hudson Bay less than 1 mile to the north.

Springs in Permafrost Areas

Data describing 17 representative springs, all situated in the area of discontinuous permafrost, are listed in Table XXV. Some of these have been described by Brandon (1965). Small springs and seeps commonly occur throughout much of the discontinuous permafrost area of Canada. They can be seen in the faces and along the toes of many of the rock bluffs bordering the valleys of such major streams as Yukon, Porcupine, and Peel Rivers and their tributaries. The larger springs commonly have

Table XXIV

Volume of Water Pumped from Mines in Northern Hydrogeological Region

Mine	Location	Mined	Production (tons of ore daily)	Major Mining Method	Depth (feet)	Water pumped (gpd)	Remarks
1. United Keno Hill Mines, Ltd. (Elsa)	Elsa, Y.T.	Ag–Pb, Zn	100	Square set	770	40,000	
2. United Keno Hill Mines, Ltd. (Hector and Calumet)	Elsa, Y.T.	Ag–Pb, Zn	350	Square set	1,300	50,000	
3. Consolidated Discovery Mines Ltd.	Yellowknife, N.W.T.	Au	150	Cut and fill	3,950	14,400	
4. Consolidated Mining and Smelting Co. Ltd.	Yellowknife, N.W.T.	Au	500	Cut and fill	3,500	470,000	
5. Giant Yellowknife Mines Ltd.	Yellowknife, N.W.T.	Au	1,000	Cut and fill	2,000	280,000	Mostly introduced water
6. Eldorado Mining and Refining Ltd.	Uranium City, Sask.	U	2,000	Cut and fill	4,000	280,000	
7. Gunnar Mines Ltd.	Gunnar, Sask.	U	2,000	Blast hole	1,980	360,000	Connected to old pit
8. International Nickel Co. of Canada	Thompson, Man.	Ni–Cu	6,000	Cut and fill	2,000	216,000	
9. Sherritt Gordon Mines Ltd.	Lynn Lake, Man.	Ni–Cu	4,250	Blast hole	2,400	250,000	After some grouting

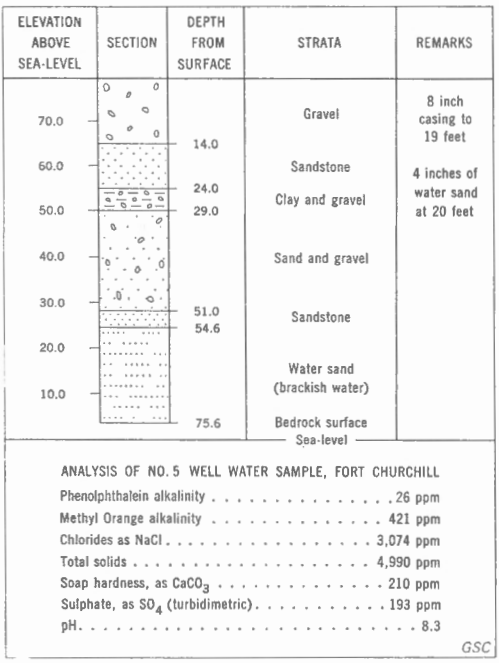


FIGURE 63.
Well No. 5 (percussion), Fort Churchill, Manitoba, 9th to 25th October, 1947 (*from* Crumlish, W. S., 1948. Exploratory well drilling in permafrost, Fort Churchill, Man.; Engineer Research and Development Laboratories, Fort Belvoir Va., Rept. No. 1045, p. 62).

their chief source in limestone. These waters evidently pass through the rock along large solution channels produced by continuous circulation of the water. Large cavities in limestone formed by groundwater are visible in the walls of some canyons along South Nahanni River. Most of these are at present dry although a few are discharging negligible quantities of water. The quantity of water yielded by gravel is probably large, but the discharge is seldom localized and it is difficult to estimate the volume.

Usually little is known regarding the relation of the springs to permafrost, and consequently it is not possible to determine if the water is supra-, intra-, or subpermafrost or a mixture of these types. Small springs or seeps issuing from gravel may result from thawing of the active layer or infiltration of precipitation down to the impervious frozen ground. The larger springs are those whose waters flow from taliks. The sulphate content of the water is frequently high. Boyle (1965) suggests that this sulphate was probably derived from the oxidation of pyritiferous schists, phyllites and quartzites.

In the Northwest Territories, several thermal springs exist in the upper part of the valley of south Nahanni River and in the area surrounding Flat and Caribou Rivers. Information has been obtained on a few of these springs by various members of the field staff of the Geological Survey of Canada. They can be identified from the air by the bright orange-red coloration of the adjacent rocks and the deposits of buff-coloured tufa associated with them. One spring has a deposit of tufa extending out from it for a distance of 1,000 feet. At one point this deposit is about 800 feet wide.

Table XXV

Chemical Analyses of Spring Water, Northern Hydrogeological Region

(*parts per million*)

Name	Takhini, Y.T.	Liard River, B.C.	Atlin, B.C. (warm)	Atlin, B.C. (cold)	Portage Brule, B.C.	Roche-qui-Trempe-à-l'Eau, N.W.T.	Clauson Creek N.W.T.	Great Slave Lake, N.W.T.		
Location	20 mi. NW of Whitehorse	Mile 497 Alaska H.	O'Donnel R. valley, s. of Atlin	O'Donnel R. valley, s. of Atlin	Left bank of Liard R. downstream from mouth of Coal R.	E. bank Mackenzie R. 2 miles downstream from Wrigley	Right bank S. Nahanni River upstream from Clauson Creek	High Point near Buffalo River	Sulphur Point	Windy Point
Date taken	July 24/62	July 12/60	July 29/62	July 29/62	Aug. 1/62	Aug. 9/60	Sept. 9/60	July 27/60	July 26/60	Sept. 9/60
Aquifer	Limestone(?)	Limestone	Limestone	Limestone	Dolomite	Limestone(?)	Limestone	Dolomite, limestone	Dolomite, limestone	Dolomite, limestone
Temperature (°F)	116	108–118	80	48	90–112	70–88	95–98	37	37	
Depth to permafrost	—	—	—	—	—	—	—	—	—	—
Appearance and odour	Few white particles	Clear	Clear	Orange precipitate	Specks of iron oxide	H_2S	H_2S			Green-brown deposit
pH	7.5	8.1	7.7	7.7	7.3	7.7	7.1	7.9	8.1	7.7
Colour	0	8	0	0	0	10	15	15	10	20
Conductivity	2,676	1,301	401	5,824	1,217	17,981	8,636	14,397	3,160	921.7
Hardness: total	1,854	738	209	4,878	627	3,229	1,650	3,077	1,877	150
non-carbonate	1,746	589	1.4	0.0	32.8	3,078	1,418	2,816	1,625	62.2
Calcium	586	234	55.1	8.8	125	925	470	713	483	41.1
Magnesium	95.1	37	17.3	1,180	76.5	223	116	315	163	11.5
Sodium	30.1	15	2.6	30.3	40.6	3,220	1,076	2,340	112	124
Potassium	7.7	9.8	0.7	9.2	34.0	34.5	75	6.9	2.9	1.5
Iron: total	0.62	tr.	0.04	7.4	0.82	0.11	0.12	0.01	0.03	0.11
Manganese	0	0.02	0	0.14	0.08	0	0.1	0	0	0
Ammonia	0	—	0	0	—					
Bicarbonate	132	182	253	5,961	725	184	283	318	307	107
Sulphate	1,721	566	11.3	60.5	77.0	2,810	732	2,550	1,577	62.3
Chloride	1.5	18.5	0.8	14.0	63.6	5,226	2,271	3,869	124	192
Fluoride	4.4	3.2	0.14	0.08	0.33	3.0	0.6	1.6	0.0	0.0
Nitrate	7.6	3.0	0	0	0	0.8	0.2	2.4	0.2	0.0
Silica	41	52	14	74	40	24	28	9.4	10	2.9
Sum of constituents	2,559	1,036	226	4,311	814	12,556	4,908	9,964	2,624	488
Flow (cfs)	50	1.5 (Aug./61)	1.0 (Aug./61)	5.0 (Aug./61)	5–10	0.2–10	40	1	1	

Table XXV (conc'd)

Chemical Analyses of Spring Water, Northern Hydrogeological Region

(*parts per million*)

Name	Flat River, N.W.T.				Old Fort Island, N.W.T.	Pelly River Y.T.	Ross River Road, Y.T.
Location	Right side of Flat R. 5 mi. downstream from McLeod Crk	Right side of Flat R. 5.5 mi. downstream from McLeod Crk	Left side of Flat R. 4.5 mi. downstream from McLeod Crk	4 mi. west of Junction of Flat and S. Nahanni Rivers	Mile 336, Mackenzie R.	Right side of Pelly R. at Hoole Canyon	Mile 158 from Watson Lake
Date taken	July 14/64	July 14/64	July 13/64	July 30/64	Aug. 8/60	Aug. 15/64	Aug. 17/64
Aquifer	Limestone	Limestone	Limestone	Limestone (?)	Limestone	Gravel	Gravel (?)
Temperature (°F)	51	63	54	39	53	34	42
Depth to permafrost	—	—	—	—	—	—	3′
Appearance and odour	Clear	Clear	Cloudy	Clear		Clear	Cloudy
pH	7.3	7.3	7.6	7.4	7.9	7.7	8.0
Colour	5.0	5.0	20	5	10	10	15
Conductivity	880	665	207	1,011	1,535	492	598
Hardness: total	512	379	99.9	628	886	273	339
non-carbonate	0	58	22.7	167	703	30	37
Calcium	151	96	31.5	197	250	64.8	83.2
Magnesium	33	34	5.2	33.0	64	27.0	31.9
Sodium	2.0	1.9	1.0	1.6	27.5	3.3	2.4
Potassium	3.5	5.2	0.7	1.1	2.0	1.9	1.4
Iron: total	0.31	0.05	0.69	3.6	0.03	2.8	tr.
Manganese	0	0.06	0	0	0	0	0
Ammonia	0	0	0	0		0.1	0.1
Bicarbonate	563	391	94.1	562	223	296	368
Sulphate	45.3	60.4	21.8	154	683	28	38
Chloride	0.2	0.3	0.1	0.4	36.9	0.4	0
Fluoride	0.48	0.38	0.13	0.55	1.0	0.15	0.21
Nitrate	0	0.1	0	0.2	0.4	1.8	0.2
Silica	22	21	5.4	6.2	6.6	9.8	11
Sum of constituents	534	410	112	671	1,180	283	350
Flow (cfs)	.1	.3	.2–25	15	40–125	.1–5	—

The temperatures of the water in the springs included in the table vary from a few degrees above freezing to 118°F. The following causes have been advanced to explain the origin of the heat in the warmer springs:

1. Friction along planes of movement in mountain-built regions (Warren, 1927).
2. Chemical action of the waters with minerals present in the rocks.
3. The absorption of heat from a high thermal area that may surround recent intrusive rocks (Brandon, 1965).
4. Radiation of residual heat from the interior of the earth (Beach, 1952). Meteoric waters passing through a deep flow system are warmed and subsequently emerge at ground surface.

Beach indicates the geothermal gradient is 1°F for every 75 feet in the Foothills of Alberta. Brandon (1965) states that some temperature logs of wells in the north show an increase of about 1 degree per 60 feet of depth. The great variation in the temperatures of the spring waters described in the tables is probably due in part to mixing with cold water near the ground surface. This would also account for the variation in the chemical composition. The waters from the cold springs at Hoole Canyon on Pelly River and along the Ross River Road are believed to be a mixture of supra- and intrapermafrost water. The aquifers are similar coarse-grained gravels overlaid with frozen organic material and silt. It has been reported that there are places in some northern rivers that seldom freeze over, even during the coldest winters. This may be the result of a warm spring discharging into the bottom of the river.

The chemical character of groundwater depends to a large extent upon the solubility of the materials through which it is passing and the length of time they are in contact. The quantity of dissolved minerals can be large if the material exposed to the water is highly soluble and the water is in contact with it for a long time. As limestone and dolomitic limestone constitute the source rocks for most bedrock springs in Canadian permafrost regions, the predominant salts in these waters are those of calcium and magnesium. Water high in chlorides has probably been associated with rocks of marine origin. Sulphates could be derived by the solution of gypsum or the oxidation of pyrite.

Groundwater Supply in Permafrost Areas

As far as is known, there are no Canadian communities in the area of continuous permafrost that are obtaining a permanent water supply from groundwater. Such settlements are usually situated along the sides of rivers and lakes from which they obtain their water supply. At Inuvik, the largest Canadian community in the area of continuous permafrost, the supply is obtained from a nearby lake. The settlements of Arctic Red River and Fort McPherson, farther south on Arctic Red and Peel Rivers respectively, obtain their water from the rivers and adjacent ponds. Unlike Inuvik, neither of these communities has a distribution system.

Brandon (1965) has described in some detail the hydrogeology at several communities in the areas of discontinuous permafrost, most of which obtain their water supply from groundwater. Table XXVI shows the more important of these communities, most of which are along the Alaska Highway. Some of the communities are

Table XXVI

Chemical Analyses of Well Water, Northern Hydrogeological Region

(*parts per million*)

Name	Alaska Highway, Yukon Territory									
Location	Watson Lake Motel, Mile 632	R.C.M.P. Teslin, Mile 804	Whitehorse Municipal Well, No. 1	Casa Loma Motel, Mile 923	R.C.M.P. Haines Junction, Mile 1016	Haines Pump Station, Mile 1026	C.N.T., Destruction Bay, Mile 1083	Motel Burwash, Mile 1095	Donjek Station, Mile 1124	Customs House, Mile 1202
Date taken	July 31/62	July 29/62	July 25/62	July 25/62	July 22/62	July 22/62	July 21/62	July 19/62	July 19/62	July 18/62
Aquifer	Sand, gravel	Sand (?)			Sand	Sand, gravel		Gravel		Sand, gravel
Temperature (°F)			39–40.5				36	32.5	34	44
Depth to permafrost							Layers of frozen ground present	Several layers of frozen ground present to depth of 90 feet		Not present
Appearance and odour	Slight grey ppt	Slight reddish ppt	Clear	Clear	Clear	Clear	Clear	A few black particles	Slight rust ppt	Clear
pH	7.7	8.0	8.0	8.3	8.4	8.2	8.2	8.0	8.0	8.0
Colour	0	0	0	0	0	0	5	0	25	5
Conductivity	346	1,042	368	470	571	331	656	508	727	482
Hardness: total	168	618	159	247	300	176	332	232	401	252
non-carbonate	5.5	250	31.0	26.6	20.4	20.1	89.4	0	0	30.6
Calcium	49.5	95.8	28.3	72.5	33.6	53.3	38.5	42.2	72.1	82.1
Magnesium	10.8	91.9	21.4	15.9	52.5	10.3	57.3	30.8	53.7	11.3
Sodium	5.8	6.3	12.2	4.8	9.9	1.6	14.2	19.3	9.9	3.7
Potassium	1.2	2.9	1.9	1.4	4.5	0.9	4.1	5.0	6.4	1.6
Iron: total	0.27	2.9	0.27	0.02	0.18	0.15	0.63	0.05	3.4	0.09
Manganese	0.04	0.14	0	0	0	0	0.02	0.3	0.2	0.03
Ammonia	0		0	0	0	0	0.2	0.0	1.0	0.1
Bicarbonate	198	449	156	268	341	189	296	288	509	270
Sulphate	7.2	249	55.9	35.6	40.2	20.4	121	40.9	1.6	22.0
Chloride	10.2	2.2	0.9	1.4	0.7	0.4	0.8	1.9	2.6	5.2
Fluoride	0.06	0.73	0.24	0.25	0.34	0.08	0.44	0.29	0.46	0.08
Nitrate	4.7	0	0.8	0.4	0.8	0.4	0.8	1.2	0.4	8.3
Silica	14	15	12	17	14	10	17	23	26	15
Sum of constituents	201	685	210	282	330	191	400	306	424	282

Table XXVI (conc'd)

Chemical Analyses of Well Water, Northern Hydrogeological Region

(*parts per million*)

Name	Yukon Territory		Northwest Territories			
Location	Dawson	Forestry Office, Mayo	Fort Simpson R.C. Mission Boiler House	Wrigley Airport	Fort Smith, Well No. 4	Hay River, Well No. 6
Date taken	July 13/62	July 15/62	Aug. 3/62	Aug. 3/60	1958	May/58
Aquifer	Sand, gravel		Silt, sand	Gravel	Gravel, fractured granite	
Temperature (°F)						
Depth to permafrost		Layers of frozen ground present	Present	Present	Present sporadically	
Appearance and odour	Clear	Clear				
pH	7.6	7.5	8.0	8.1		
Colour	5	5	15	15		
Conductivity	204	190	1,063	553.7		
Hardness: total	96.5	98.1	464	308	755	280
non-carbonate	27.4	17.2	154	18.9		
Calcium	25.4	26.3	113	79.2		
Magnesium	8.0	7.9	440	26.7		
Sodium	2.1	1.5	57.0	1.6		
Potassium	0.8	0.7	2.3	2.5		
Iron: total	0.02	0.14	0.02	0.04	0.3	2
Manganese	0.01	0.01	0	0		
Ammonia	0.1	0.1				
Bicarbonate	84.2	98.4	378	352		
Sulphate	29.8	17.0	164	19.9	396	13
Chloride	1.5	0.4	62.9	0.8	186	23
Fluoride	0.14	0.08	0	0		
Nitrate	1.4	1.5	16	5.0	0	0
Silica	6.2	4.8	5.9	7.7		
Sum of constituents	117	109	652	317	1,262	422

situated beside a large stream or lake and obtain their water from wells dug or drilled into unfrozen material close to the surface water. Dawson, the farthest north, obtains its water from two wells excavated into alluvial sand and gravel. The wells are situated along the north side of Klondike River near its junction with Yukon River. Wells at the town of Hay River, situated on Vale Island in Hay River, obtain their water from alluvial silt, sand and fine-grained gravel overlying lacustrine clay and clay till. A well-drilling program conducted in 1958 indicated that the underlying shale bedrock was tight and the most likely place to obtain a groundwater supply for the town was in the alluvial material. Permafrost was present in some of the holes. A representative log of one of these holes is as follows:

Hole No. 6

Depth (feet)	*Description of Material*
0– 1	clay
1– 8	fine red beach sand
8–25	coarser grey sand
25–30	clay, sand seams
30–45	soft clay
45–56	soft gravel and clay
56–62	clay

Static water level 3 feet from surface.

Before the construction of a filtration plant and distribution system, the community of Fort Smith on Slave River was dependent upon wells for its water supply.

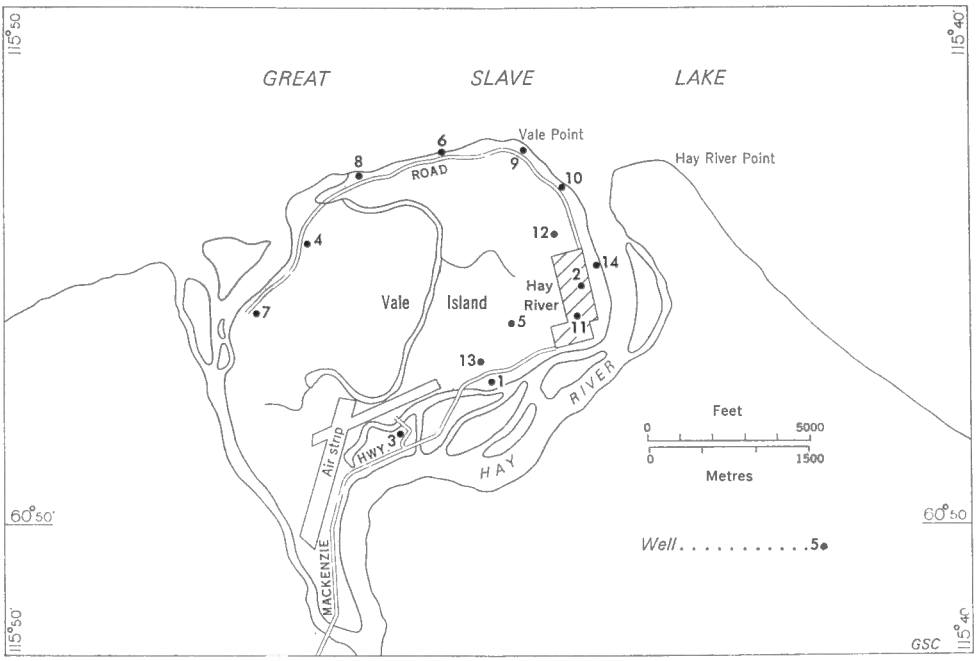


FIGURE 64. Location of test wells drilled, 1958, Hay River, Northwest Territories.

The wells were dug 19 to 36 feet into sand overlying clay. In 1958 a well-drilling program was carried out at Fort Smith to determine whether there was enough groundwater available to supply the community (Fig. 65). The results were negative and the testing was abandoned. The following is the log of a representative hole put down during the test-drilling program.

Hole No. 4

Depth (feet)	*Description of Material*
0– 10	sandy brown clay
10– 35	sand
35– 45	silty clay
45– 50	silty, blue clay
50–173	soft, blue clay
173–180	granite wash

Depth to water 56 feet.
Test pumping was carried out from the granite.
Yield was 40 gpm for 88-foot drawdown.

The water supply for Whitehorse, the capital of Yukon Territory, is obtained from both Yukon River and a small well field. The latter consists of three wells situated on the east side of Yukon River about 3,000 feet upstream from the bridge. The yield of the wells varies from 200 to 320 gpm, the water being obtained from coarse-grained alluvium. One reason for using groundwater is to warm the river water during cold periods, the temperature of the groundwater being higher than

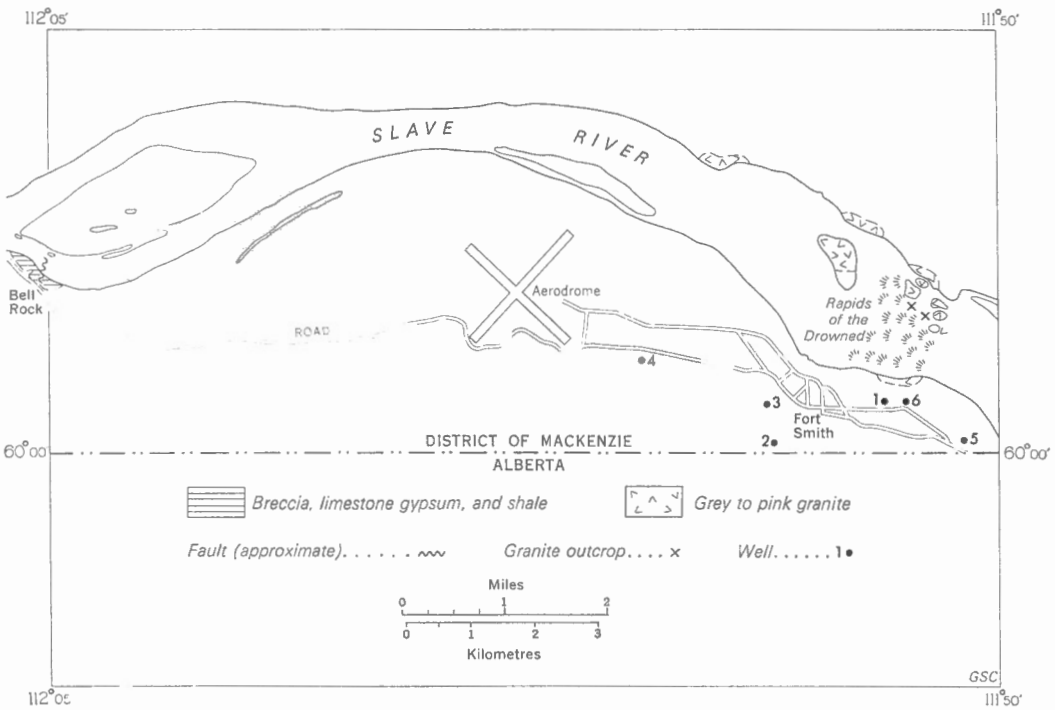


FIGURE 65. Location of test wells drilled, 1958, Fort Smith, Northwest Territories.

that of the river water during the winter. As far as is known, frozen ground was not encountered in any of the three wells. The following is a log of one of the wells at Whitehorse.

Well No. 1

Depth (feet)	*Description of Material*
0– 3	topsoil
3–19	dirty gravel
19–30	clean coarse gravel
30–40	coarse gravel, silty

Capacity 320 gpm
Specific capacity 64 gpm/foot
Static water level 6 feet

Between the Alaska boundary and Mile 550, the Alaska Highway lies within the area of discontinuous permafrost. For much of this distance the highway traverses broad almost level valleys filled with varying thicknesses of glacial deposits. Similarly situated are the Whitehorse–Mayo–Dawson road, the Canol road and the Ross River–Watson Lake road. Rivers such as the Rancheria, Yukon, Dezadeash, and White drain the valleys in which the Alaska Highway lies. In the few areas where the gradient is steep these rivers are relatively straight and fast, but usually they are braided streams meandering across the valley floors and leaving behind old meander scars and dry, abandoned channels. Numerous lakes occur in these valleys, the largest being Teslin Lake and Kluane Lake, which are the source of the two rivers of the same name.

Commencing at ground surface, a general log of the glacial material filling these valleys is: alluvium—glacio-lacustrine silt and/or clay—glacio-fluvial sand and gravel —till—bedrock.

One or more of these materials could be absent in any one locality. Irregular lenses of sand and gravel frequently occur within the till and sometimes separate the till and underlying bedrock.

In general there is no shortage of groundwater, the chief aquifer being the coarse-grained, glacio-fluvial material that is frequently capable of yielding large quantities of water to properly developed drilled wells. Coarse-grained alluvium is also an important aquifer that can be tapped by shallow, dug wells. This water, however, is often perched and only a small, temporary supply is available. The groundwater in the valleys can occur as supra-, intra-, or subpermafrost water or a mixture of any of the three, depending upon the presence of permafrost and its relation to the water-bearing material.

Unfortunately there is little information concerning the occurrence of permafrost in wells drilled along the Alaska Highway. This is especially true for wells drilled at the various army camps during highway construction in 1943 and 1944. Several wells in the community of Watson Lake, ranging in depth from 30 to 70 feet, obtain excellent supplies of water from coarse-grained, glacio-fluvial gravels. Permafrost, if present here, apparently has little effect upon the supply of groundwater. At Teslin, shallow wells drilled near the lake obtain small but adequate supplies, possibly from thin sand lenses in glacio-lacustrine silts.

Several wells put down along the Alaska Highway between Whitehorse and the Alaska boundary produce varying quantities of water, chiefly from glacio-fluvial gravels. Permafrost is sporadic in this area but becomes more common as the boundary is approached.

Most communities along the Whitehorse–Dawson highway are situated at river crossings. Supplies of groundwater adequate for local needs are produced by shallow wells dug or drilled into coarse-grained alluvium. Permafrost is usually not present in these wells, although it has been observed close to ground surface in test pits put down along the road a short distance back from the streams. The most satisfactory supplies of groundwater in the community of Mayo are from wells some distance from Stewart River. These wells obtain their water from coarse-grained, glacio-fluvial gravel. The alluvial deposits close to the river are not a good source of water. Permafrost is uncommon throughout the community, although it has been reported in some localities.

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Chapter IX

AQUIFER EXPLORATION AND DEVELOPMENT

by A. Lissey

Introduction

Aquifer exploration followed by aquifer development precedes the production of groundwater. Exploration consists in selecting an aquifer; development refers to preparing it for exploitation. The purpose of this chapter is to (1) describe the techniques used in exploration, (2) describe well construction as a part of development, and (3) present a plan of exploration and development for the water-supply engineer to follow.

The first part of this chapter outlines a suggested procedure for exploration; the second, describes the techniques used; the third, outlines a suggested procedure for development and describes the steps of well construction.

The guide or check list is to aid in choosing and following the best sequence of operations for various conditions. The formulae, techniques, and equipment referred to are all presented and illustrated in detail in the references cited. For smaller supplies the same procedure is recommended but on a scale appropriate to the requirements. For example, a procedure for an individual home supply can be summarized as follows:

1. Ask the neighbours
2. Ask drillers with local experience
3. Ask groundwater agencies (Table II)
4. Employ a reputable driller.

Exploration Procedure

Sound exploration enables the engineer to select the most suitable aquifer to develop so that he can alleviate his water-supply problem in the most economical way. The exploration procedure suggested consists of five steps.

The first step is to define the area to be explored—usually it should be circular. Its centre should be the point of utilization, and its radius should be equal to the distance from this point to the nearest usable source of surface water. The second step is to find all aquifers occurring within the exploration area. The first part of this

step is a preliminary geological study to outline those parts of the area where aquifers are likely to occur. Existing data such as regional geology maps, aerial photographs, topographic maps, soils maps, and well logs are used in this study. The last part of this step is the use of geophysical techniques in favourable areas to find buried aquifers from the surface. The third step in exploration is to determine the size, shape, and subsurface attitude of each aquifer found. This is done partly by surface geophysical techniques and partly by subsurface techniques that require drilling. The fourth step is to determine the hydraulic head, hydrochemistry, and permeability of every aquifer. The head is determined by measuring water levels in wells and relating these to mean sea-level. The hydrochemistry is determined by analyzing water samples from springs and wells in the aquifer. The permeability is determined either in the laboratory or in the field. The fifth and last step is to select, by comparative elimination, the aquifer that has the most desirable hydrochemistry, highest permeability, greatest head, largest dimensions, shallowest burial depth, and shortest distance from the utilization point. This aquifer will be the most capable of yielding the desired quality and quantity of water for the least cost.

At this point the engineer is prepared to estimate and compare the cost of utilizing surface water and the cost of utilizing groundwater. If the latter is more economical, he must develop the selected aquifer.

Exploration Techniques

Surface Geophysical Techniques

The water-supply engineer requires a sound knowledge of surface geophysical techniques because they are the most important means of finding aquifers. For this reason the commonly used geophysical techniques are described in detail. A short account of those techniques that may be used in the future is presented at the end of this section.

Geophysics measures the physical properties of rocks. An aquifer is detected in the subsurface because its properties differ from those of the surrounding rock. The properties usually measured in aquifer exploration are electrical resistivity and elasticity. The respective techniques used to measure these properties from the surface are earth resistivity and seismic surveys.

Earth Resistivity Surveys

An earth resistivity survey consists in measuring the voltage drop between two potential electrodes on the earth's surface when an electrical current is passed through a volume of earth material between a pair of current electrodes. The apparent resistivity is determined by the ratio of the measured voltage to the impressed current. It is also a function of electrode separation. It depends on the true rock resistivities and the distribution of the rocks within the volume of earth through which the current effectively passes. The true resistivity of a rock varies with lithology, porosity, permeability, water content and quality, and temperature. The volume of earth through

which the current effectively passes varies with the current electrode spacing and the resistivities of the rocks conducting the current.

Resistivity equipment may be operated on either alternating or direct current. Low-frequency alternating current eliminates the problem of contact potentials at the electrodes interfering with field measurements. Although direct-current equipment is subject to polarizing potentials at the electrodes, it can attain greater depths of current penetration than can equivalent alternating-current equipment. To avoid the effects of electrode contact potentials with direct-current equipment, porous pots containing a copper electrode immersed in a saturated solution of copper sulphate can be used as potential electrodes.

Positioning electrodes may be placed in various geometric arrangements called electrode configurations. The arrangements are of the Wenner (Wenner, 1916), the Schlumberger (Stefanescu and Schlumberger, 1930), and the Lee partitioning (Lee, *and others*, 1929) types. An adaptation of these symmetrical arrangements, commonly called the "three-electrode" configuration, is often made by theoretically placing one of the current electrodes at infinity. A particular configuration is selected either to simplify field procedure or to use a form of expression for resistivity that will simplify interpretation of the results.

There are two distinct ways in which each configuration can be used in the field. In one method, termed "lateral profiling", the distance between adjacent electrodes in the configuration is kept constant while the entire array is moved between stations and a single determination is made at each station. In the other method, termed "vertical profiling", the midpoint between potential electrodes remains fixed and the current electrode spacing is varied for each reading. Either lateral or vertical profiling is used to find an aquifer. Once the aquifer is found, the lateral extent of an aquifer having a uniform depth of burial can be determined by lateral profiling. The depth to the top and, in some cases, the thickness of an aquifer can be determined by vertical profiling.

Both lateral and vertical profiling data can be interpreted qualitatively as to the presence or absence of aquifers in any area. The quantitative interpretation of vertical profiling data as to depths and thicknesses of aquifers is based on the concept of a layered earth. In this concept, each rock type having different physical properties occurs in vertically successive, laterally infinite, physically isotropic and homogeneous, horizontal layers. Theoretical quantitative interpretations of vertical profiling data can be made by an analysis for the two-layer case (Tagg, 1934), a type-curve matching method for two-, three-, and four-layer cases (Mooney and Wetzel, 1956; La Compagnie Générale de Géophysique, 1955), or a linear approximation method for multiple layers (Unz, 1962). Empirical methods of quantitatively interpreting these data have been described by Gish and Rooney (1925), Moore (1945), and Barnes (1952).

Often, however, none of the available methods give valid interpretations (Hobson *and others*, 1962, p. 1077). This may be due to one or more of the following sources of error (Lennox, 1962): (1) dipping strata, thickness variation, and irregularities of surface topography depart from the assumption of horizontal layering; (2) depositional heterogeneity, such as that characteristic of glacial till, and physical unhomogeneity characterized by gradual directional changes of resistivity within one rock type

invalidate the assumption of physically homogeneous layers; and (3) electrical anisotropy characterized by higher conductivity in one direction than in another does not meet the assumption of physically isotropic layers. In Hobson's reply to Lennox (1962, p. 293), the additional problem of decreased instrument sensitivity due to a dry surface layer was suggested as a further handicap to reliable interpretation of resistivity results.

The amount of error introduced into quantitative interpretation by these effects is not known. Until it can be determined, resistivity surveys should not be conducted to determine aquifer depths and thicknesses without complementary subsurface data to provide geologic control.

From work done in Alberta, Lennox (1962) concludes that present earth resistivity surveys for aquifer location are best suited to thick sand and gravel aquifers near surface and become less suited as aquifer thickness decreases or overburden thickness increases. Besides being a tool for aquifer exploration, the earth resistivity survey is a reliable means of detecting subsurface salt-water boundaries (Swartz, 1939). In this regard it may become an important tool of groundwater management where salt-water intrusion threatens producing aquifers.

Seismic Surveys

A seismic survey consists in producing elastic waves by a sudden release of energy at a point near the surface and detecting their arrival times at various other points on the surface after they have travelled through subsurface rocks. Energy may be released on or below the surface by mechanical impact or by explosive detonation. The velocity of seismic waves through any rock in the earth's crust depends on the elastic properties of that rock. These properties are affected by lithology, porosity, and the degree and nature of fluid content. Seismic-wave propagation through these rocks adheres to the optical laws of reflection and refraction at interfaces where abrupt velocity changes occur.

There are two types of seismic surveys. The seismic-reflection method is based on analyzing the travel paths of reflected seismic waves. The seismic-refraction method is based on analyzing both travel paths and velocities of seismic waves that have been refracted at rock interfaces. The reflection method is used to determine geological structure in areas where the seismic velocities of the various formations are known. The refraction method is used to determine both the structure and the seismic velocities of formations. It is better suited to shallow work than the reflection method.

For aquifer location, only the seismic-refraction method is used at present, although high-resolution reflection surveys could be adopted. These refraction surveys are either of the conventional type, in which multitrace recording equipment is used and energy is released by explosive detonation, or of the less expensive "hammer" type, in which a single geophone is used and energy is released by the mechanical impact of a sledge hammer on a steel plate.

The interpretation of seismic-refraction data as to aquifer depths and thicknesses is based on the assumption that each rock type occurs as vertically successive, laterally infinite, physically isotropic and homogeneous, horizontal layers, each having a greater

seismic velocity than the immediately overlying one. Analyses of refraction data are started by plotting observed times of first seismic-wave arrival at a geophone against the geophone distance from the energy source. For the simple two-layer case, this graph consists of two linear segments. The slope of the first segment is inversely proportional to the seismic velocity of the upper layer because the arrival of the direct wave occurred before the arrival of refracted waves at small distances from the energy source. The slope of the second segment is inversely proportional to the seismic velocity of the lower layer because, beyond a critical distance from the energy source, refracted waves arrive before the direct wave. By calculating the seismic velocities of the two layers and using either the critical distance or the intercept time, the thickness of the upper layer can be calculated. The critical distance is determined on the graph from the intersection of the two segments. The intercept time is determined by the intersection of the extrapolated second segment and the time axis. Similar methods are used for cases involving more than two layers. In all cases standard corrections for buried energy sources and for irregular surface topography are made where applicable. Formulas for these calculations are given by Jakosky (1957) and Dobrin (1960).

The following problems (Lennox, 1962) complicate the application of seismic-refraction surveys to aquifer location: (1) the existence of velocity reversals invalidates the assumption of increased velocities with depth; (2) the heterogeneous character of glacial drift exhibited by sand and gravel lenses within till, and continuously increasing velocities with depth characteristic of some rocks, invalidate the assumption of physically homogeneous layers; (3) seismic anisotropy, generally characterized by greater horizontal velocities than vertical velocities, does not meet the assumption of physically isotropic layers. In addition, the "blind zone" problem, which occurs with thin layers such as might be encountered near the lateral limits of an aquifer, invalidates depth determinations of lower layers. Although the problems of dipping beds, weathered surface layers, low velocity contrasts, ground-coupled air waves, and weak first-cycle energies can be handled, their recognition in any seismic-refraction survey is essential. Many of these problems can be recognized and corrected only with the aid of independent subsurface logging data. Without these subsurface data the value of much seismic work also remains doubtful.

The seismic-refraction method of aquifer location is best suited to shallow aquifers where high velocity contrasts occur. The depth of penetration limits the applicability of the hammer seismic method to very shallow occurrences whereas the problem of geophone spreads limits the applicability of the conventional refraction method to slightly greater depths. Where low velocity contrasts or deeply buried aquifers occur, the high-resolution seismic-reflection method, complemented by continuous sonic logs or by well-shooting programs in drill-holes, is more applicable.

Future Techniques

Of the other surface geophysical techniques available, the induced polarization method shows the greatest promise for future application to aquifer location. This method uses a standard earth resistivity technique to pass current through subsurface

material but is based on measuring the rate of decay of the potential induced in this material after the current is arrested rather than the potentials associated with the current during its application. This method has been applied to groundwater exploration by Vacquier, *and others* (1957), who concluded that the induced polarization method can be used to detect the depth to the water-table and might be used to estimate aquifer permeability.

Another surface geophysical method that shows promise for future aquifer prospecting is the gravity survey. This method detects near-surface changes in density by measuring variations in the earth's gravitational pull on a known mass at different positions on the earth's surface. As indicated by Lennox (1962) and Hall and Hajnal (1961), gravity surveys may provide a cheaper alternative than either high-resolution seismic-reflection methods or earth resistivity methods in finding aquifers when they are deeply buried or where low velocity contrasts occur.

Subsurface Techniques

The engineer must have a sound knowledge of the subsurface techniques used in aquifer exploration because they accurately determine the physical relations by which aquifers are compared. All subsurface techniques require the sinking of holes into the aquifer. This section gives a detailed description of how holes are made. It presents a brief account of the logging techniques by which these holes can be used in aquifer description and a brief account of the pumping-test methods by which an aquifer's shape and size are determined indirectly.

The procedure for making a hole in any material consists of two operations: (1) disintegration of the material and (2) removal of the disintegrated material. In boring, these two operations are accomplished by a single process; in drilling they are accomplished by separate, independent processes. When applied to holes in earth materials, the disintegration operation can be divided into percussive-crushing and rotational-abrasion processes, while the removal operation can be divided into mechanical, hydraulic, and pneumatic-discharge processes. The two last-mentioned processes can be subdivided into standard and reverse-circulation types. With standard circulation, the fluid is discharged from the hole via the annulus between the wall of the hole and the drill rods. With reverse circulation the fluid is discharged via the hollow rods themselves.

The drilling methods used to make holes in earth material are classified in Table XXVII according to the kind of removal and disintegration processes used.

The boring methods used to make holes in earth materials are based solely on high-energy forms of removal processes. Jetting is a boring method that utilizes the standard circulation hydraulic removal process. All other boring methods utilize a mechanical removal process. These methods include digging, augering, driving, and boiling-down. Digging and augering are accomplished either by hand or by power machinery. Power augers are divided into the excavator-bucket, continuous-flight, and discontinuous-flight types.

Boring methods are rarely used in groundwater exploration because, in order to keep the total energy requirements within the operational limits of the particular

process employed, they must be restricted to material that is easily disintegrated. Speed of operation and reliable sampling are requisites that limit the remaining methods of obtaining subsurface aquifer information via holes to cable-tool drilling and standard rotary drilling.

Table XXVII

Drilling Methods

			DISINTEGRATION PROCESS	
			Percussive Crushing	Rotational Abrasion
REMOVAL PROCESS	Pneumatic	Reverse Circulation		
		Standard Circulation	Mining Air Hammer	Air Rotary
	Hydraulic	Reverse Circulation		Reverse Rotary
		Standard Circulation	Modified Mining Air Hammer	Standard Rotary Turbine Rotary Diamond Rotary
	Mechanical		Cable Tool	Chilled Shot Core Drilling

Cable-tool Drilling

A cable-tool rig consists essentially of (1) a steel frame on which are mounted a power unit, three hoisting drums, and an adjustable-stroke walking beam and (2) a telescoping derrick with three sheaves at the top. Power is transmitted to the drums and walking beam via a countershaft driven by Vee belts. The unit may be skid-, trailer-, or truck-mounted for portability.

One hoisting drum, termed the bull reel, spools the drilling line. The free end of this line passes around the end of the walking beam and over the crown sheave at the top of the derrick, to be attached to a string of drilling tools that operates down the hole. The drill string, as run in the hole, consists of a bit, a drill stem, a set of jars, and a rope socket. The drill stem is used to weight the bit. The jars, consisting of two connecting links, are used to loosen any tools that become lodged in the hole. The rope socket is used to attach the drilling line to the drill string and to rotate the

string after each impact of the bit. The percussive disintegration operation of the drilling procedure is accomplished by a vertical rising and falling of the bit as the drilling line is in effect shortened and lengthened respectively. This shortening and lengthening is caused by the reciprocating motion of the walking beam, which is imparted by a double-crank-arm drive from the countershaft.

A second hoisting drum, termed the sand reel, spools the sand line. The free end of this line passes over the sand line sheave at the top of the derrick and is attached to either a dart-valve bailer or a sand pump by means of a quick-change bailer link. Both the bailer and the sand pump consist of a short length of tubing with a one-way valve at the bottom, and both are used to remove the disintegrated material from the hole. A sand pump is additionally equipped with a plunger that draws material into the tool more forcefully than a bailer. Material drawn into a bailer or sand pump is hoisted out of the hole and deposited at the surface.

Except for a small amount of water, added to slurry the cuttings for ease in bailing, and except for formation fluids, a hole drilled by the cable-tool method is essentially void of any material. For this reason holes are maintained in unconsolidated sediments with casing that prevents caving. This casing is installed so that the bottom of the casing string is always kept near the bottom of the hole. The hole is made to progress by alternately drilling the material and driving the casing for short distances each time. The casing lengths, which are riveted, welded, or threaded together, are driven by means of drive clamps attached to the drill stem. The impact of the clamps on the drive head of the top casing length forces the casing down the hole. A drive shoe prevents damage to the bottom of the casing string.

A third hoisting drum, termed the casing reel, is used to raise lengths of casing for this procedure. This drum spools the casing line, the free end of which passes over the casing-line sheave on top of the derrick and is attached to casing clamps.

The standard sampling procedure used in cable-tool drilling is to retain some of the material removed from the bottom of the hole by bailing. This method provides accurate samples, particularly where the hole is progressively cased. Cable-tool rigs are ideally set up to obtain undisturbed formation samples of unconsolidated material by the split-barrel drive-core method (Anon., 1960a). Besides the problem of caving, the greatest difficulty encountered in cable-tool sampling is that of preventing saturated unconsolidated sediments from heaving. Heaving can best be described as a "blow-through" or "sand boil" (Ferris, *and others*, 1962, p. 80) into the hole. It can be prevented only by maintaining a hydrostatic head inside the hole that is greater than the natural hydrostatic head of the heaving formation at the bottom of the hole. This is done with a column of drilling fluid.

The cable-tool drilling method provides the most reliable means of making a hole in any type of aquifer. The greatest drawback to the extensive use of this method is the slow rate of drilling—less than 10 feet an hour. As a tool of aquifer exploration, however, cable-tool drilling is used extensively in shallow consolidated aquifers that occur at depths of less than 100 feet. This use is justified because at shallow depths cable-tool drilling rates can exceed rotary drilling rates and because cable-tool bit life exceeds rotary bit life in hard materials.

Standard Rotary Drilling

The basic standard rotary drilling rig consists of (1) a steel frame on which are mounted two hoisting drums, a revolving table, a double-acting piston pump, and either a mechanical or a hydraulic pull down and (2) a hydraulically raised derrick with two sheaves at the top. The unit may be skid-, trailer-, track-, or truck-mounted for portability. Power is supplied either by power take-off from the vehicle's motor or by a separate motor mounted on the drill frame.

One hoisting drum, called the main reel, spools the drilling line. The free end of this line passes over the main crown pulley and either is attached directly to the swivel bail or, after passing through a travelling block that hooks the bail, is attached to an anchor eye on the underside of the crown block. A stationary pulley called a snatch block is sometimes attached to the anchor eye, and the drilling line passes over this pulley to be dead-ended at the base of the derrick. On large rigs an elevator connection replaces the bail, and the drilling line is attached to quick-release clamps called elevators, again either directly or, more commonly, via a travelling block. With this last-mentioned arrangement an additional permanent crown pulley is preferred to the snatch block. The purpose of the drilling line is to support the drill string, which, as run in the hole, consists of a bit, a number of drill collars, a number of drill pipes, and, at the upper end, a kelly. The kelly is a length of drill pipe with a special outer contour such that it must rotate with the rotary table and yet is still free to move vertically through the table. The upper end of the kelly is attached to the swivel, the housing of which remains stationary even while the kelly is rotating. All connections within the drill string are by threaded pin and box joints.

A second hoisting drum, called the sand reel, spools the sand line, the free end of which passes over the sand line sheave and is usually attached to a swivel socket for subsequent threading to various tools. On rigs not equipped with elevators one of these tools is a clevis that fits into a hoisting-plug bail. The plug is used to pick up each stand of drill pipe to be added to the drill string as the hole deepens. During such connections the drill string is suspended in the hole by the top tool joint of the uppermost stand of drill pipe on slips placed in the table while the disconnected kelly is suspended in the derrick on the drilling line. After the connection is made, the slips are removed and the drill string is lowered until the top tool joint of the added stand is caught and suspended in the slips. The kelly is then attached and drilling proceeds, deepening the hole to the point where the swivel is resting on the table. The kelly is then removed and stood aside, and the string is again suspended by the top tool joint on the slips while a new stand of drill pipe is added. This procedure is repeated stand after stand until the final depth of the hole is attained. On rigs equipped with elevators the disconnected kelly is placed in a rathole—a shallow hole adjacent to the main hole—and the elevators are used to pick up additional stands.

Disintegration of the material by rotational abrasion is accomplished by rotating the table and hence the entire drill string including bit. Weight is maintained on the bit by the drill collars. Standard procedure is to keep the collars and bit in compression and the drill pipe in tension. For exceptionally hard rock at shallow depths the smaller rigs utilize pulldowns to maintain the entire string in compression. Two basic

kinds of bits are used—drag bits and rock bits. Drag bits are used in soft unconsolidated material. Rock bits, which may be of the cone or roller type, are used in hard consolidated rocks.

The process of removing the disintegrated material from the hole is the standard hydraulic circulation of drilling mud. Mud circulation starts from the pumps, moves up a standpipe alongside the derrick, through a flexible kelly hose, through the swivel, down the inside of the kelly, drill pipe, and collars, and out openings in the bit. From here it returns up the hole through the annulus into a settling pit, which overspills into a suction pit at ground surface, then back to the pumps via a flexible suction hose. The disintegrated material accumulates in the settling pit.

Because the hole is always full of drilling fluid, caving and heaving of unconsolidated material can be prevented by controlling mud viscosity and density. Numerous drilling-fluid additives are used to vary these two properties.

The standard sampling procedure is to retain some of the material discharged from the hole. Reliable samples are obtained when the hole is circulated clean of cuttings before the sample interval is drilled. Rotary rigs are also equipped to obtain undisturbed formation samples either by standard rotary-cone barrel in consolidated rock or by Shelby tube (Anon., 1960b) in unconsolidated material. By raising and dropping the drill string, rotary rigs can also obtain samples by the split-barrel drive-core method. Rotary drive-coring is less efficient than cable-tool drive-coring.

Standard rotary drilling provides the fastest and hence the most economical means of drilling in the exploration of all deep aquifers. It is especially suited for drilling in unconsolidated aquifers because the hole does not need casing to stand open.

Well Logging

By drilling and logging enough holes the size, shape, and subsurface attitude of every aquifer can be accurately determined. Well logging refers to systematically recording the changes in rock properties with depth of hole.

The properties recorded by well logs are sample description, rate of penetration, degree of caving, temperature, electrical resistivity, electrical conductivity, spontaneous electrical potential, natural radioactivity, effect of induced radioactivity, and sonic velocity. Measurements of these properties respectively furnish sample logs, drilling-time logs, caliper logs, temperature logs, normal and lateral resistivity logs, induction logs, self potential logs, gamma ray logs, neutron logs, and sonic logs. Sampling procedures have been described in previous sections. Sample logs are discussed by Low (1958). Drilling-time logs are discussed by Kirby (1954) and by Shepherd (1958). Caliper logging is described by Tapper (1958) and by the Schlumberger Well Surveying Corporation (Anon., 1958a). The latter also presents detailed accounts of the remaining geophysical logging methods.

At present, only sample, normal-resistivity, and self-potential logs receive widespread use in aquifer exploration. In the future gamma ray and neutron logs will be used extensively because they can be run in cased or open holes with or without drilling fluid. Thus, besides providing means of logging both cable-tool and rotary holes on

the same basis, they permit old wells and abandoned holes, about which no original data are available, to be used as additional control points.

Logs are used qualitatively to differentiate lithologies, define formation boundaries vertically, and provide a means of subsurface correlation. In the future the quantitative use of logs may become extensive. Self-potential logs may be used to determine water qualities, and neutron logs may be used to determine aquifer porosities.

Aquifer-limits Tests

Certain kinds of pumping tests, called aquifer-limits tests, can be used to determine the lateral extent of an aquifer. These tests are used to hydrologically confirm and adjust the mapping of aquifer boundaries previously based on the results of geophysics and drilling.

Aquifer-limits tests are based on the fact that the cone of depression around a pumped well becomes distorted when it encounters a zone of lower permeability. These tests are run in existing wells and temporary test wells near aquifer boundaries.

One aquifer-limits test is based on the theory of images (Ferris, 1949). This method treats the lateral limits of an aquifer as an impermeable barrier. The time at which the cone of depression starts to distort is determined from a semi-logarithmic plot of the drawdowns in an observation well. This time can be related to the distance between observation and image wells. If data from three or more observation wells are used, the image well and hence the aquifer boundary can be found.

Another pumping-test method originally developed to determine the size of oil reservoirs can be adapted as an aquifer-limits test (Nind, *in preparation*). This method attempts to determine the minimum and maximum distances between a pumped well and the aquifer's lateral boundaries. It does not require observation well data. The minimum distance is determined from the time defining the upper limit of transient flow when the cone of depression starts to distort. As in the image theory, this time is indicated by a break in the original slope of semi-logarithmically plotted time-drawdown data. The maximum distance is determined from the time marking the start of semi-steady-state flow, when the cone, having encountered the farthest boundary, ceases to distort. This time is indicated by the point where arithmetically plotted time-drawdown data assume a straight-line relation.

Determination of Permeability

Before the engineer can choose the best aquifer to develop he must determine the permeability variation in each. This variation is found by mapping individual permeability determinations.

The means of obtaining permeability determinations can be divided into field and laboratory methods. The former are generally more reliable.

Laboratory Methods

The cheapest though least reliable laboratory method of determining permeability depends upon a mechanical analysis of an aquifer sample. Formulas that attempt to relate permeability to measurable physical properties are applied to this analysis.

These formulas vary from a simple relation with grain size only (Hazen, 1893) to a complex relation between porosity, degree of sorting, grain shape, density, and grain size (Fair and Hatch, 1933). This method can be used for unconsolidated aquifers only.

Under a more reliable method, which utilizes a permeameter, fluid is passed through an aquifer sample and the rate of flow and resulting head drop are measured. Knowing the sample dimensions, the rate of fluid flow, and the head loss, one can determine the permeability of the sample. Permeameters may be of the discharging or non-discharging type. The former can operate under either a constant head or a falling head. Permeameter techniques have been described by Wenzel (1942).

There are two disadvantages in using laboratory methods for permeability determination. One is that the samples used may not be representative of the aquifer from which they are taken. This is especially true of unconsolidated aquifers where the original grain packing is disturbed by sampling. The other disadvantage is that they cannot be applied meaningly to samples of aquifers in which fracture permeability is developed.

Field Methods

The most common method of determining aquifer permeabilities in the field is with a pumping test. Analyses of pumping tests are based on solutions of the continuity equation, which describes the induced flow of a slightly compressible fluid through a homogeneous, isotropic medium.

Theis (1935) presented a non-equilibrium solution of the continuity equation for an infinite, horizontal aquifer bounded above and below by impermeable material in which exists horizontal, radial, laminar flow towards a line sink that discharges at a constant rate. In describing the drawdown at any point in a very large, flat-lying, confined aquifer that is vertically bounded by impermeable strata, this solution is applicable if flow is produced by a single fully penetrating well. For constant-rate pumping tests (Bruin and Hudson, 1955), where such conditions exist, this solution can be applied by using either the type-curve method of test-data superposition (Wenzel, 1942) or the modified non-equilibrium method (Cooper and Jacob, 1946). The former method involves logarithmic plots of time vs. drawdown whereas the latter, an approximation, involves semi-logarithmic plots of either time vs. drawdown or distance vs. drawdown. The permeability of aquifers, where no leakage from confining beds occurs, can be determined by applying one of these methods to early time data that have not been influenced by lateral boundaries.

Hantush and Jacob (1955) derived a non-equilibrium solution of the continuity equation that includes the Theis solution as a limiting case and supplements it by treating vertically confining strata as having varying degrees of permeability through which water enters the aquifer within the zone of depression at rates proportional to the drawdowns therein. This solution is applicable in describing the drawdown due to pumping a well in large "leaky" aquifers. A graphical method of test-data superposition on logarithmic type curves (Walton, 1960) is used for this application. Again, early time-drawdown data must be used to determine the permeability of such aquifers.

A steady-state, distance-drawdown type-curve method (Jacob, 1946) can be applied to late drawdown data when the rate of cone expansion has been stabilized owing to the addition of leakage water and only if no lateral boundary conditions affect the cone.

Boulton (1963) derived a non-equilibrium solution to the continuity equation, that takes into account both the delayed yield from storage and the departure from the Dupuit assumption of horizontal flow to describe the drawdown anywhere in an unconfined aquifer bounded below by impermeable strata. This solution indicates that the Theis equation, corrected for aquifer dewatering (Jacob, 1944), can be applied to very early and to late time-drawdown data but that water released from storage between early and late times does not instantaneously accompany a head decline. In Boulton's analysis, time-drawdown data obtained during this period of delayed yield adhere to the Hantush and Jacob equation for leaky aquifers. This solution can be applied to pumping tests of wells in unconfined aquifers according to a graphical method of test-data superposition on logarithmic type curves presented by Boulton to determine aquifer permeabilities.

The field methods of determining permeability by pumping tests, which have been discussed in the last three paragraphs, were originally developed for application to data from observation wells only. Where no observation wells exist, aquifer permeability can be estimated from the initial specific capacity of pumping wells (Theis *and others*, 1954) if corrections for partial penetration (Muskat, 1949) and for well efficiency (Walton 1962, Nind, *in press*) are applied. This method is applicable only in non-leaky confined aquifers.

Field permeabilities can also be determined from pumping tests on infiltration galleries if the previously mentioned methods of well hydraulics are applied to distant observation wells (Mikels and Klaer, 1956). It is believed, however, that if observation wells are in the immediate vicinity of the gallery, adaptations of the horizontal tube-drain formulas (Kirkham, 1949) are more applicable. Permeability determinations involving the use of galleries without observation wells have not been developed.

Apart from the use of pumping tests, permeabilities can be determined in the field by using bailer and slug tests. Bailer tests can be analyzed in various ways (Skibitzke, 1958; Kirkham, 1945; Hvorslev, 1951). Slug tests can be analyzed by a method proposed by Ferris and Knowles (1954). Permeability determinations by application of horizontal drain formulas can also be done in the field (Van Hoorn, 1960; Ferris, 1962).

Development

Procedure

After the engineer has completed the five exploration steps and decided to produce groundwater, he must develop the best aquifer in the most economical way. The development procedure suggested comprises three steps. The first consists in determining how the aquifer is to be exploited; the second in choosing the facilities needed to exploit it that way. The final step consists in constructing these facilities.

Production Policy

The way in which an aquifer is to be exploited is determined by one of three possible production policies. The policy chosen must be in accord with the laws of water-resource management. One policy is that production from an aquifer will not exceed its safe-yield allotment. Another is that production will exceed the aquifer's safe-yield allotment by an artificially recharged increment. The third policy is that production from an aquifer will exceed the safe-yield allotment in such a way that groundwater storage is depleted, or mined, at a predetermined rate.

The choice of policies depends on the quantity of water required and on the aquifer's safe-yield allotment. The quantity of water required varies with each supply problem. It must be predicted by the water-supply engineer. The safe-yield allotment is assigned by basin management. It is not determined by the local engineer. The engineer, however, must have an understanding of the concept of safe yield and the methods of determining it in order to choose a production policy acceptable to water-resource management.

Concept of Safe Yield

Safe yield as defined by Todd (1959) is "the amount of water which can be withdrawn from . . . [a groundwater basin] annually without producing an undesired result." Disregarding the quality, and economic and legal factors, safe yield is that part of natural groundwater recharge to a basin that can be withdrawn without depleting groundwater storage. An aquifer's safe-yield allotment is a part of the enclosing basin's safe yield.

The total safe yield of a basin can best be defined by considering the hydrologic budget equations of a basin. By using the law of mass conservation and short duration of time, the following "general hydrologic budget" equation can be applied to any small drainage basin:

$$P+(SI)=(ET)+(R_s)\pm(\triangle S)\pm(\triangle S_{sm})+GW \qquad (1)$$

where P=precipitation
SI=surface water inflow via channel or pipeline diversions
ET=surface evaporation and soil evapotranspiration
R_s=surface run-off
$\triangle S$=change in surface storage (+=increase, −=decrease)
$\triangle S_{sm}$=change in soil moisture storage (+=increase, −=decrease)
GW=groundwater recharge.

The associated "groundwater hydrologic budget" equation is:

$$(GW)=(GO-GI)+(ET_g)+(R_g)\pm(\triangle S_g)+(Pr) \qquad (2)$$

where GO=groundwater outflow
GI=groundwater inflow
ET_g=groundwater evapotranspiration
R_g=groundwater run-off (baseflow)
$\triangle S_g$=change in groundwater storage (+=increase, −=decrease)
Pr=groundwater withdrawal.

In an undeveloped basin where only natural conditions exist, SI and Pr equal zero.

For a fixed value of GW and zero change in S_g, the maximum value of Pr in equation 2 is achieved when R_g, ET_g, and GO are reduced to zero. Under such conditions Pr=GW+GI. The interaction, however, between the right-hand terms of equation (2) may be such that any positive value of Pr reduces S_g without affecting the values of the other terms. Thus the safe yield of any small basin is some value between zero and GW+GI.

It should be noted that Pr can exceed GW without affecting $\triangle S_g$ only if GI > GO, *i.e.* if "infiltration to deep aquifers" (Kudelin, 1958) occurs within the basin. Considering, however, a larger flow system, of which the small basin in question is only a part, GI and GO both equal zero and the maximum value of safe yield for the larger basin cannot exceed the natural groundwater recharge to it. The magnitude of this larger system may be at the continental scale (Meyboom, 1963).

Groundwater withdrawal can exceed safe yield and still not decrease groundwater storage if the value of GW can be increased artificially. Artificial recharge may be effected either by decreasing the right-hand terms or by increasing the left-hand terms of equation 1. Increase of the left-hand terms involves artificial rain-making and surface-water diversions. Although the right-hand terms interact and cannot be decreased independently, most artificial recharge projects attempt to reduce R_s or $\triangle S$.

Determination of Safe Yield

A reliable method of predicting the safe yield prior to production is the practical long-range aim of groundwater research. Empirical methods, by which safe yield is evaluated after a lengthy period of production, cannot be used when original production policy is being established.

One method of safe-yield prediction is based on a procedure for determining future water levels in wells from a pumping test (Bruin and Hudson, 1955). Another is based on a flow-net analysis of the piezometric surface across an aquifer (Bennett and Meyer, 1952). Neither method is considered to give reliable predictions of safe yield. They are employed for lack of better practical methods.

Theoretically the ideal method of determining safe yield is to evaluate all terms in equations 1 and 2 independently and to determine the terms' interaction. Budget analyses have been used in a few basins to evaluate these terms (Rasmussen and Andreasen, 1959; Schicht and Walton, 1961), and electric analog computers (Skibitzke, 1961) could determine the interaction between terms. The suggested combination of these two techniques should provide a practical method of reliably determining the safe yield of a basin.

Choice of Facilities

In choosing the facilities needed to exploit the selected aquifer, the engineer must aim at the lowest cost-benefit ratio, adhere to the production policy chosen, and bear in mind the geologic and hydrologic limitations of the aquifer. Cost is determined by the amount and type of the materials and services required for the construction, operation, and maintenance of the facility. Benefit is determined by the product of the

quantity and value of the water produced. Although influenced by the purpose for which it is to be used, the value of water is governed by the law of supply and demand.

There are four major types of facilities. These are extraction units, replenishment units, transmission devices, and treatment plants.

Extraction units can be classified according to the attitudes of the intake and delivery sections (Table XXVIII).

Table XXVIII

Types of Extraction Units

		Delivery Section	
		Lateral	Vertical
Intake Section	Vertical	Water mines	Vertical wells
	Lateral	Kanats	Galleries

The intake section refers to that part where water from the aquifer enters the unit. The delivery section is that part of the unit between the intake section and the surface of the earth.

The best known water mine exists near Frankfurt, Germany (Nöring, 1955). Kanats served ancient cities in the Middle East (Tolman, 1937). Water mines and kanats are not used extensively in modern times. Galleries are used to induce infiltration through unconfined aquifers from hydraulically connected surface-water sources (Kazmann, 1948; Rorabough, 1951). Vertical wells are the most extensively used extraction devices.

Replenishment units can be divided into water spreads, recharge pits, and injection wells. Water-spreading techniques consist of flooding methods, basin methods, ditch methods, natural channel methods, and irrigation methods (Todd, 1959).

Transmission devices consist of simple or composite conduits arranged in single or multiple lines. Simple conduits may be either gravity drains or pressure pipelines. Gravity drains are characterized by flow with zero pressure head. Flow in pipelines is under positive pressure head. Transmission devices are described by Davis (1952), King (1954), and Hobbs (1950).

Treatment plants vary from simple chlorination units to complex iron-removal or desalinization units. The type of treatment required depends on the difference between the quality of water available and that desired. The quality of water desired depends on the purpose for which it is to be used. The three main uses are domestic, industrial, and irrigation supply. Respective quality tolerances are listed by the

American Water Works Association (Anon., 1946), the New England Water Works Association (Anon., 1940), and Wilcox (1948 and 1955). One advantage of groundwater is that it rarely requires treatment for undesirable organic material and does not need to be filtered. These are additional problems common to surface-water treatment.

To produce groundwater, extraction units and transmission devices are always required. Replenishment units are required only if artificial recharge is necessary. Treatment plants may or may not be required.

Construction of Vertical Wells

Location

The last and most important step in development is the construction of the chosen facilities. This section considers the construction of vertical wells only.

The first phase of well construction is locating the site where maximum yield can be attained. The potential yield of a well is the product of the available drawdown and the specific capacity.

To obtain maximum specific capacity, a well should be located (1) as far as possible from barrier boundaries such as the aquifer's lateral limits, (2) as close as possible to recharge boundaries such as hydraulically connected surface bodies of water, (3) in the thickest part of the aquifer, and (4) in the most permeable part of the aquifer.

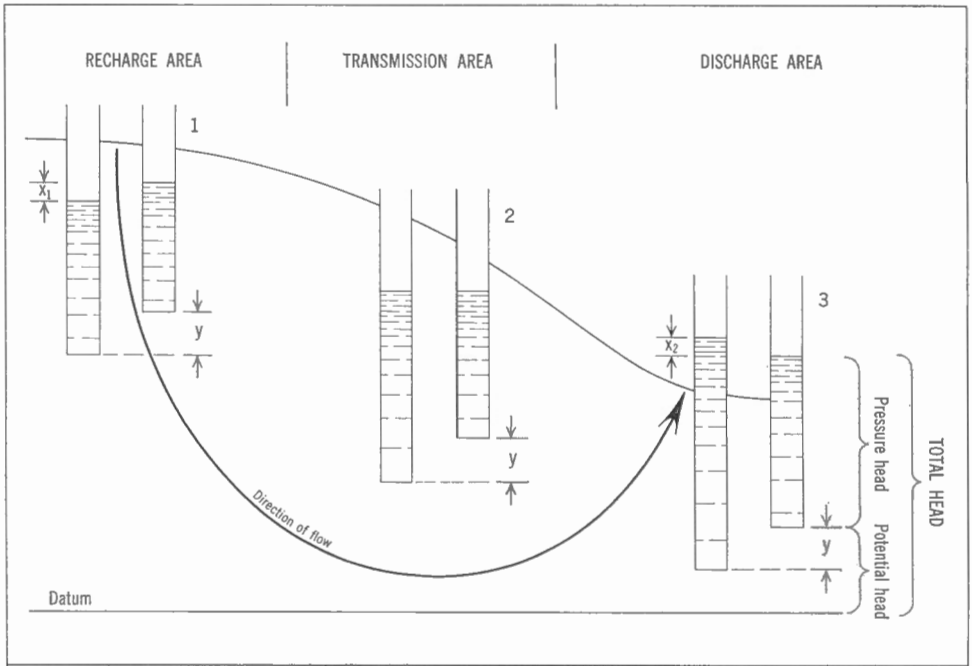
To obtain maximum available drawdown a well should be located where the greatest natural pressure head occurs. In general, natural pressure heads increase everywhere with depth. The increase is least in recharge areas, greater in transmission areas, and greatest in discharge areas (Fig. 66). Thus, a well should be located in the discharging portion of the aquifer and should be completed at the bottom of the aquifer.

Rarely will an aquifer contain one particular site where all the desirable characteristics of maximum yield are combined. Usually a number of equally favourable locations will be found by compromise. The factor of shortest transmission distance then determines which location should be chosen.

Design

After locating a site, the engineer enters the design phase of well construction. A well's intake and delivery sections are designed independently. The economic aim of well design is maximum efficiency.

In designing a well's intake section the engineer must determine (1) the type (screened or open-hole), (2) cross-sectional shape and size, and (3) length. For maximum efficiency the length of intake equals the thickness of aquifer (Muskat, 1949). The size of intake depends on the pumping rate and the type depends on whether the aquifer is consolidated or unconsolidated. In the screened intakes the engineer must also determine (1) the mode of screening (natural-strainer or gravel-wall), (2) the size and variation of screen openings, and (3) the type of material



	PIEZOMETER NEST		
	1	2	3
Increase in depth (Decrease in potential head)	y	y	y
Direction of natural flow	Down (total head decreases with depth)	Laterally (total head constant with depth)	Up (total head increases with depth)
Arbitrary change in total head	x_1	0	x_2
Increase in pressure head in deeper piezometer	$y - x_1$	y	$y + x_2$
Rate of pressure head increase	$\frac{y - x_1}{y}$ ft/ft	1 ft/ft	$\frac{y + x_2}{y}$ ft/ft

GSC

FIGURE 66. Head relationships in natural flow system (partly *after* Hubbert, 1940).

constituting the screen. When artificial gravel walls are required, he must also determine the composition and size variation of the gravel constituting the wall. These all depend on aquifer grain size.

In designing a well's delivery section the engineer must determine (1) the type (cased or uncased), (2) cross-sectional shape and size, and (3) length. Cased wells are used where the overburden is unconsolidated material. The shape and size are determined by the pumping equipment to be installed, and the length is fixed by the burial depth of the aquifer. For cased delivery sections he must determine (1) the mode of casing (single string, stovepipe (or telescopic), or multiple string), (2) the manner of coupling (threaded, welded, or riveted), (3) the type of material constituting casings, and (4) the type of corrosion control (coated, cemented, or cathodic). These depend largely on the method of installation to be used.

Installation

After a well is designed, it must be installed at the chosen site. Because most wells are of the cased-and-screened type, this is the only type discussed in this section.

There are two methods of installing casings and screens in wells. In one, the hole is drilled or bored first and the casing or screen is positioned later. In the other, the casing or screen is installed as the hole is made.

Screens are contemporaneously installed by driving, bailing-down, or hydraulic jetting. In the first method, screens with conical points, called drivepoints, are forcibly projected into the aquifer by mechanical impact. In the second method, screens equipped with a bail-down shoe at the bottom are used. A bailing pipe that fits inside the screen and extends to surface is temporarily threaded to this shoe. The screen is lowered into position by operating a bailer inside the bailing pipe. Material below the screen is removed by bailing and the screen slowly drops into the hole below. In the third method, a screen equipped with a wash-down check valve at the bottom is used. A wash pipe that extends inside the screen to surface is temporarily threaded to this valve. Water directed through the wash pipe jets a hole into which the screen drops.

Screen positioning, after a hole has been made to accommodate it, is done by either the pull-back or the open-hole installation technique. In the pull-back technique the casing is temporarily placed at the bottom of the hole in the aquifer; the screen is then lowered within the casing to its desired position; and finally the casing is raised, the screen being thereby exposed to the aquifer. In the open-hole technique the unprotected screen is projected through the stationary casing into the hole below.

Casings are contemporaneously installed by driving. This is done either by mechanical impact or by the application of constant pressure with hydraulic cylinders. For small-diameter shallow wells such as those used for dewatering systems the casing and screen are driven as an integral unit. Casing installation after the construction of a hole is done simply by lowering the complete string to the required depth. Casing may also be installed by alternate driving and drilling as described in the section on cable-tool drilling.

The construction of cased and screened wells that require an artificial gravel pack around the screen is usually accomplished with the aid of an outer casing within which a permanent inner casing is used. This outer casing may or may not be left as a permanent fixture after the pack is placed. Bennison (1947) details a number of construction methods used to install gravel packs, but only four of these are used at present. These are called the temporary casing, open-hole, bail-down, and stabilized methods.

To install the screen by the temporary-casing method of gravel-pack installation, the pull-back technique is used with the outer casing. Before the casing is raised, however, the annulus between the centred screen and the casing is filled with the required pack material.

In the open-hole method, the screen is centred in an underreamed or oversize hole by either the standard pull-back technique with inner casing or the open-hole technique of screen installation. The pack material is placed between the screen and the wall of the hole.

In both the temporary-casing and the open-hole methods of installing gravel packs the material is placed from the bottom up via small-diameter pipes in the annulus between the inner and the outer casings. As the pack material is positioned these pipes are raised.

In the bail-down method of gravel-pack installation, the pack material, which is stocked in the space between the inner and the outer casings, settles from the top down alongside the screen as it is bailed into place. In the stabilization method the screen is set by any of the forementioned methods, and pack material, which has been placed in the annulus between the inner and the outer casings, settles around the screen as the finer fractions of rock material constituting the aquifer are removed from the vicinity of the screen by well-stimulation techniques.

During the installation of cased-and-screened wells, provision is made for sealing the open top of the screen from the aquifer. This is usually done by swedging a lead packer to fit between the bottom of the innermost casing and the top of the screen. Rubber packers may be used in place of lead packers. In some cases light casing, which may or may not extend to surface within the inner casing, is attached to the top of the screen.

Stimulation

The final phase of well construction is well stimulation, or well development. This is the process by which aquifer permeability in the immediate vicinity of a well's intake section is increased artificially. The increase is accomplished by removing rock material. Stimulation increases well efficiency.

Direct rock removal consists in drawing the fine-grained particles through the screen and bailing or flushing them out of the well. The process of drawing these particles into the well is accomplished by creating turbulent flow in the aquifer toward the well with sufficient velocity to transport this material in suspension. During this process a number of small grains may bridge screen openings and prevent entry of further material. By reversing the direction of flow these "sand bridges" are disrupted. During such development, flow into or out of the screen is effected by rapid, large-differential, pressure-head adjustments within the well.

The standard methods used for well stimulation are surging by solid surge plungers or valved surged plungers, backwashing with compressed air, surging with compressed air, overpumping with compressed air, rawhiding with a turbine pump, hydraulic backwashing with bailing, and hydraulic backwashing under pressure (Anon., 1941).

More recent techniques of well stimulation are high-pressure jet cleaning (Anon., 1958b), dry ice development (Bennison, 1947), acid injection (Kingston, 1947), and explosive vibration (Mylander, 1952). Oil-well techniques such as hydraulic sand-fracturing (Clark, 1949) and nitro-shooting (Rison, 1929) may be used to develop consolidated aquifers in the future.

The degree of improved efficiency by well stimulation is determined by multi-rate pumping tests. These are described by Walton (1962), Russell (1963), and Nind (*in press*).

Summary

The proposed five-step plan of exploration consists in (1) defining the area for exploration, (2) finding the aquifers in that area, (3) describing them geologically, (4) describing them hydrologically, and (5) choosing the best of them. This plan enables the engineer to decide how best to alleviate his water-supply problem.

If groundwater utilization is required, the proposed three-step plan of development enables the engineer to exploit the selected aquifer in the most economical way conformable with water-resource management. These steps consist in (1) choosing a production policy, (2) choosing the facilities needed, and (3) constructing the extraction, transmission, recharge and treatment facilities as required.

The most important aspects of exploration are surface geophysics, subsurface drilling, and permeability determination. The most important aspect of development is well construction.

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