

OPEN FILE REPORT #85-28

Study of the Hydrogeology of
the White Lake Basin British Columbia

by M. B. Grant and F. A. Michel

Earth Physics Branch
Energy, Mines and Resources Canada
1 Observatory Crescent
Ottawa, Ontario
K1A 0Y3

pages : 63 & 1 foldout

Price : \$16.31

This document was produced
by scanning the original publication.

Ce document est le produit d'une
numérisation par balayage
de la publication originale.

STUDY OF THE HYDROGEOLOGY OF THE WHITE LAKE BASIN, BRITISH COLUMBIA

Abstract

Groundwaters sampled from diamond drill holes, domestic wells, springs and surface water bodies in the White Lake Basin, British Columbia, were analysed for chemical content and for oxygen-18, deuterium and tritium. Surface waters and shallow flow systems in overburden and bedrock aquifers were found to be chemically and isotopically distinct from deeper waters in the basin. Local and intermediate length flow systems are thought to exist, but there is no evidence for the existence of a regional flow system in the basin. Flow of the deep waters is influenced locally by geological and hydrological conditions.

There is no evidence that the waters from the basin have been subjected to temperatures significantly higher than measured at the discharge points

Résumé

Des échantillons d'eaux souterraines prélevés de trous de forage au diamant, de puits domestiques, de sources et de plans d'eau en surface dans le bassin du lac White (Colombie-Britannique) ont été analysés pour déterminer la teneur en produits chimiques et en oxygène-18, deutérium et tritium. Il a été constaté que les eaux de surface et les systèmes d'écoulement peu profond dans les aquifères des terrains de couverture et de la roche en place étaient chimiquement et isotopiquement distincts des eaux plus profondes du bassin. Nous croyons qu'il y existe des systèmes d'écoulement local et intermédiaire, mais nous n'avons relevé aucun indice qui appuie l'existence d'un système d'écoulement régional dans le bassin. L'écoulement des eaux profondes subit l'influence locale des conditions géologiques et hydrologiques.

Rien n'indique que les eaux du bassin aient subi des températures sensiblement plus élevées que celles constatées aux exutoires.

STUDY OF THE HYDROGEOLOGY
OF THE WHITE LAKE BASIN,
BRITISH COLUMBIA

Contract Serial No. OSB83-00223

FINAL REPORT

Prepared for:

Department of Energy, Mines and Resources

By:

M. B. Grant

F. A. Michel

March, 1983

ABSTRACT

Groundwaters sampled from diamond drill holes, domestic wells, springs and surface water bodies in the White Lake Basin, British Columbia, were analysed for chemical content and for oxygen-18, deuterium and tritium. On the basis of these analyses, three groundwater systems have been delineated. Surface waters and shallow flow systems in overburden and bedrock aquifers were found to be chemically and isotopically distinct from deeper waters in the basin.

Local and intermediate length flow systems are thought to exist, discharging into lakes and streams or as springs. There is, however, no evidence for the existence of a regional flow system in the basin. Flow of the deep waters is influenced locally by geological and hydrological conditions.

Several lakes in the area experience high evaporation due to the warm, arid climate and lack of good surface drainage.

There is no evidence that the waters from the basin have been subjected to temperatures significantly higher than measured at the discharge points. The SiO_2 geothermometer yields temperatures in the range of 0-20°C, while the Na-K-Ca geothermometer produces unrealistic temperature estimates.

ACKNOWLEDGEMENTS

The authors would like to acknowledge Dr. T. Lewis for his advice and support of this study. Thanks should also go to Robert Drimmie, for his assistance in the field, and to his staff at the University of Waterloo, Department of Earth Sciences Laboratories for their chemical and isotope analyses of samples from the study area. We would also like to thank Elizabeth Lambton for typing this report. This study was funded by the Earth Physics Branch of Energy, Mines and Resources.

TABLE OF CONTENTS

	Page
Abstract	i
Acknowledgements	ii
Table of Contents	iii
List of Tables	iv
List of Figures	v
Chapter 1: INTRODUCTION	1
1.1 Purpose	1
1.2 Location and Access	1
1.3 Previous Work	3
Chapter 2: GENERAL GEOLOGY AND HYDROSTRATIGRAPHY	4
2.1 General Geology	4
2.2 Faulting in the Area	8
2.3 Hydrostratigraphy	9
Chapter 3: INTRODUCTION TO CHEMISTRY, ISOTOPES AND GEOTHERMOMETRY	12
3.1 Water Chemistry	12
3.1.1 Chemistry of Water in Crystalline Rocks	12
3.1.2 Chemistry of Water in Glacial Deposits	13
3.1.3 Anion Evolution	14
3.1.4 Cation Exchange	15
3.1.5 Carbon Dioxide in Groundwater	16
3.2 Geothermometry	17
3.2.1 The Na-K-Ca Geothermometer	17
3.2.2 The Silica Geothermometer	19
3.3 Tritium in Groundwater	19
3.4 ^2H and ^{18}O in Groundwater	21
Chapter 4: DISCUSSION AND INTERPRETATION	23
4.1 Field Program	23
4.2 Discussion of Results	27
4.2.1 Tritium	27
4.2.2 ^2H and ^{18}O Isotopes	34
4.2.3 Water Chemistry	39
4.2.4 Geothermometry	47
4.3 Interpretations	48
Chapter 5: CONCLUSIONS AND RECOMMENDATIONS	52
References	54

LIST OF TABLES

	Page
Table 1: Field Measurements	25-26
Table 2: Isotope Data of Water Samples	30-32
Table 3: Chemical Composition of Selected Water Samples	41-42

LIST OF FIGURES

	Page
Figure 1: Location map of the White Lake Basin in South-central British Columbia.	2
Figure 2: Geology of the Penticton Tertiary Outlier.	5
Figure 3: Distribution of known and inferred major faults in the White Lake Basin.	6
Figure 4: Location of sampling points.	24
Figure 5: Mean annual tritium concentrations in precipitation and groundwaters at Ottawa, 1952-1980.	29
Figure 6: Oxygen and hydrogen isotopic compositions of waters of the White Lake Basin, B.C.	35
Figure 7: Major ion chemistry of waters of the White Lake Basin, B.C.	43

Chapter 1: INTRODUCTION

1.1 Purpose of this Study

The purpose of this study is to further distinguish groundwater systems in the White Lake Basin as originally outlined by Michel and Fritz in 1982. It will attempt to determine if local, intermediate and/or regional flow systems exist and to determine the probable origins, ages and geothermal potential of the waters.

1.2 Location and Access

The White Lake Basin is located approximately 240 kilometres due east of Vancouver and 18 kilometres south of Penticton in the lower Okanagan Valley, British Columbia (Fig. 1). It is centred on latitude $49^{\circ}18'N$ and longitude $119^{\circ}38'W$ (NTS map sheet 82 E/5).

The central portion of the basin consists of a small valley, containing White Lake, at an elevation of 580 metres above sea level that is completely surrounded by a series of mountains ranging in elevation from 915 m in the east to 1375 m in the north and south.

Regionally, highways 3A and 97 serve the area through valleys adjacent to the mountains. Secondary roads and a local network of farm and logging roads and foot paths give access to the sampling sites.

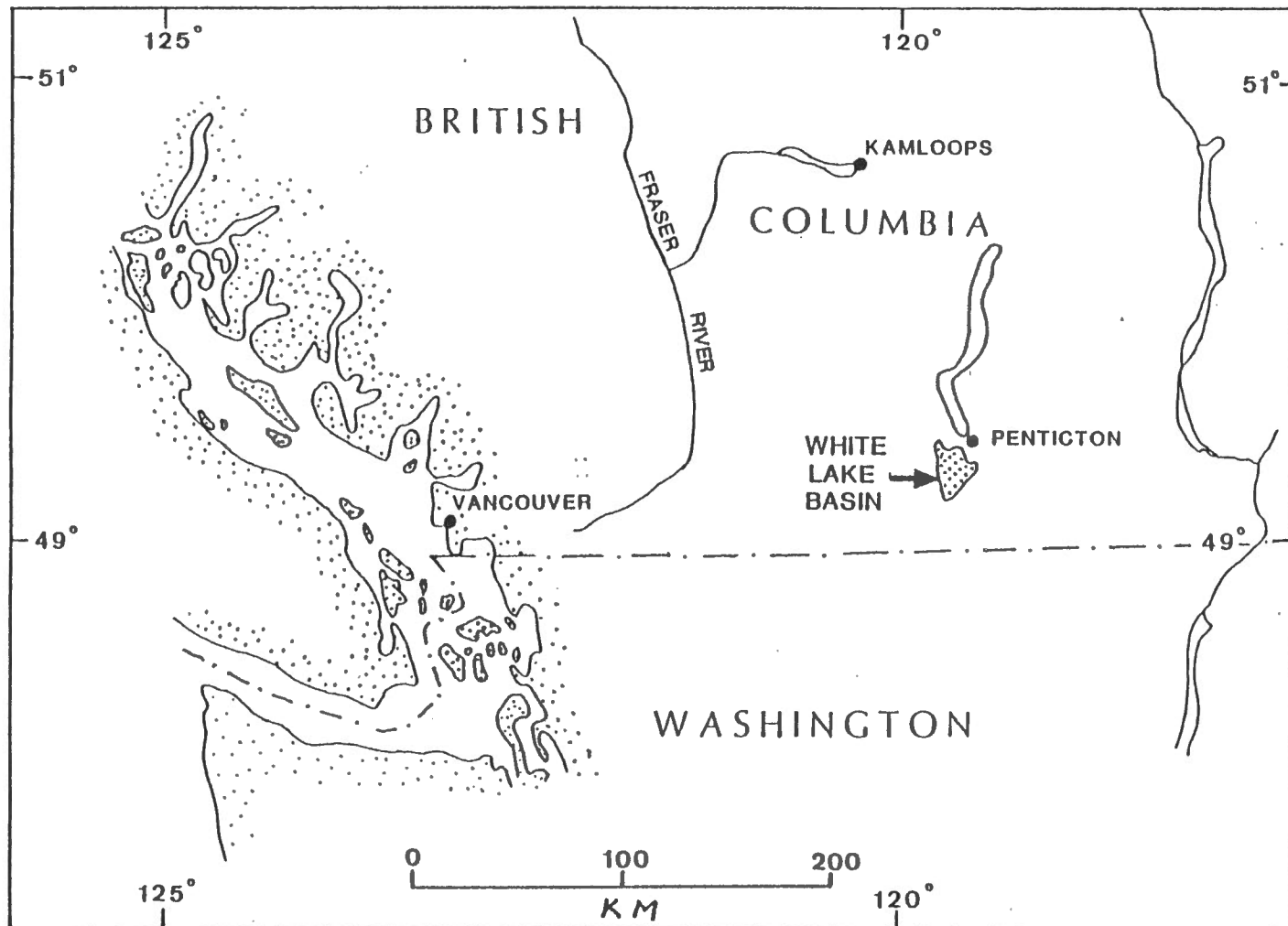


Figure 1: Location map of the White Lake Basin in South-central British Columbia.

1.3 Previous Work

Several reports concerning the study area have been written. These include a detailed geological report and initial geothermal and hydrological studies.

A detailed geological report of the White Lake area was published by Church in 1973. A preliminary geological map and cross-section was also published by Church in 1979. These reports have been used as a basis for the geological description of the area for most subsequent papers.

Jessop and Judge (1971) measured heat flow at five locations in Canada, one of which was in the White Lake Basin. They found the Penticton borehole (P-Well in this study) to have an anomalously high heat flow.

Lewis (1984) has examined the geothermal potential of the White Lake area using thermal gradient logs. On the basis of these thermal data he has suggested a northwesterly trending regional groundwater flow system which is recharged in the vicinity of White Lake.

A preliminary hydrological study of the area was conducted by Michel and Fritz (1982). They suggested the presence of three separate water masses on the basis of the geochemical and isotopic contents of the water. They also suggested that if a regional flow system exists, it probably flows in an easterly direction.

Chapter 2: GENERAL GEOLOGY AND HYDROSTRATIGRAPHY

2.1 General Geology

This summary of geology is taken from Church (1973, 1979). The rocks of the White Lake Basin are composed primarily of Tertiary volcanics and sediments known as the Penticton Tertiary Outlier (Fig. 2).

The pre-Tertiary basement is exposed at several places near the margins of the basin and consists of Triassic or older metasedimentary and metavolcanic rocks. To the south and west, the pre-Tertiary rocks are extensively intruded by Cretaceous and Jurassic granites, granodiorites and syenites.

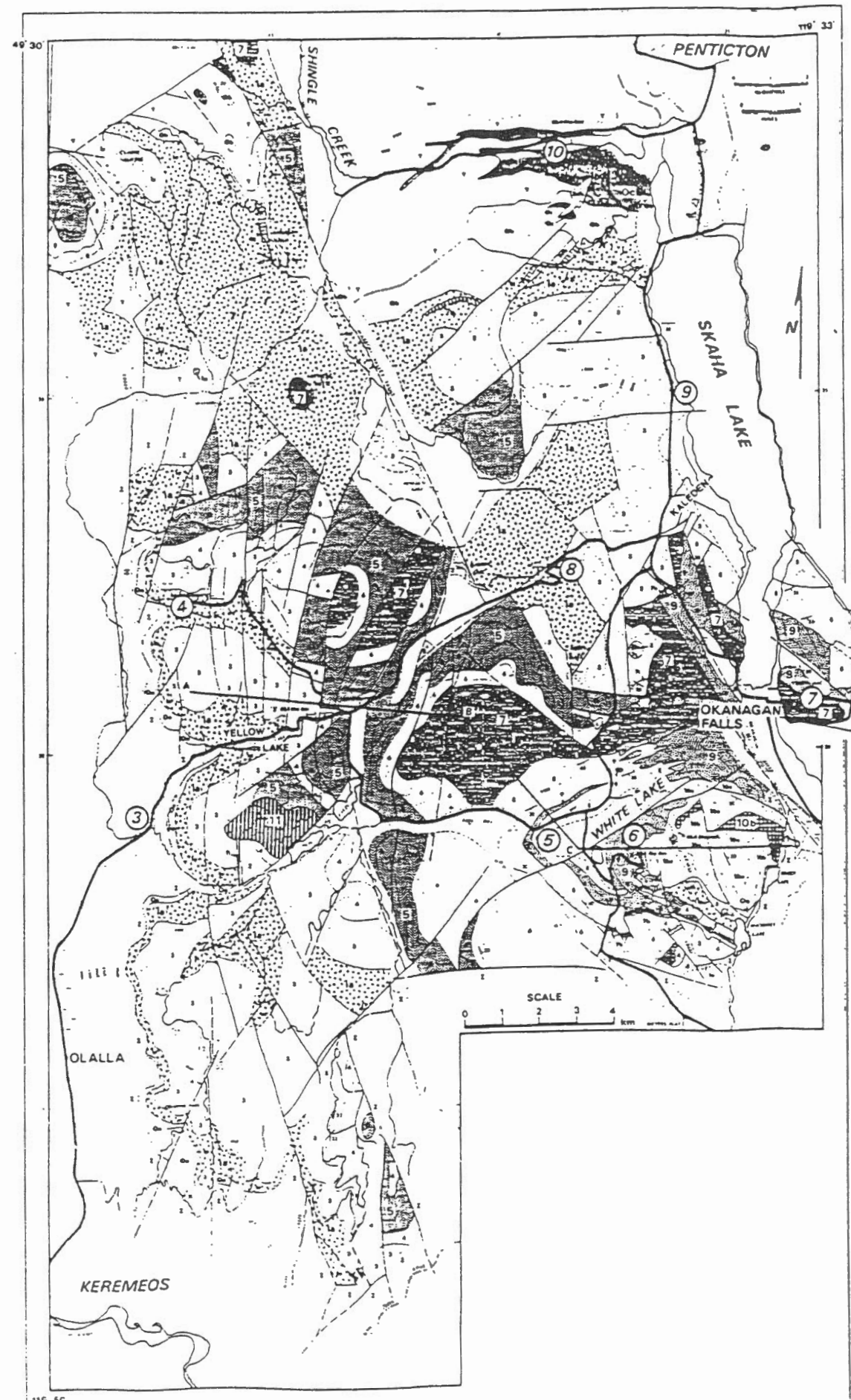
The Tertiary rocks, which dip gently to the east, are extensively cut by gravity faults striking primarily north-south (Fig. 3). The Tertiary pile is thickest and structurally lowest near the Okanagan Valley where beds reach a thickness of 2400 m, with basal beds 1680 m below sea level northeast of White Lake. Minor folds are locally developed and are believed to be the result of simple flexures in the basement rocks. Folds are best developed where the Tertiary deposits are thickest.

Church (1973) considered the early Tertiary rocks of the White Lake area to be of Eocene age. He subdivided them into the following stratigraphic formations, from youngest to oldest:

Figure 2
GEOLOGY OF THE PENTICTON TERTIARY OUTLIER

BY B. N. CHURCH

LEGEND



MIOCENE

(OLALLA RHYOLITE)

- 11 MOSTLY RHYOLITE BRECCIA, SOME MASSIVE OBSIDIAN, AND ASSOCIATED DYKES

Eocene

PENTICTON GROUP

SKAHA FORMATION

- 10a MOSTLY CHERT AND GREENSTONE SLIDE BRECCIA AND SOME TEPHRITE LAVA OVERLAIN BY POLYMIC TIC FAN-GLOMERATE

- 10b CHANNEL DEPOSIT OF GRANITE BOULDER CONGLOMERATE AND BRECCIA AND ARKOSIC SANDSTONES

WHITE LAKE FORMATION

- 9 MOSTLY VOLCANIC BRECCIAS INCLUDING PYROCLASTIC ROCKS AND LAHARS, MINOR TRACHYTIC AND ANDESITIC LAVAS

- 8 VOLCANIC CONGLOMERATE, SANDSTONES, AND SHALES

MARAMA FORMATION

- 7a AENEAS BUTTE FELDSPATHIC DACITE

- 7b MASSIVE APHANITIC DACITE LAVA AND SOME BRECCIA FORMING MOSTLY REMNANTS OF VOLCANIC DOMES

- 7c VOLCANIC CONGLOMERATE WITH CLASTS FROM THE MARRON FORMATION

MARRON FORMATION

- 6 PARK RILL MEMBER: MEROCRYSTALLINE ANDESITE LAVA AND MINOR BRECCIA

- 5 NIMPIT LAKE MEMBER: TAN TRACHYTE AND TRACHY-ANDESITE LAVA AND MINOR BRECCIA

- 4 KEARNS CREEK MEMBER: VESICULAR PYROXENE-RICH BASALTIC ANDESITE LAVA

Eocene (CONTINUED)

MARRON FORMATION (CONTINUED)

- 3 KITLEY LAKE MEMBER: TRACHYANDESITE LAVA WITH CONSPICUOUS GLOMEROPHENOCRYSTIC CLOTS OF FELDSPAR

- 2 SHATFORD CREEK MEMBER: LOCAL DEPOSIT OF BROWN ANDESITE LAVA AND BRECCIA WITH SOME QUARTZ-FILLED AMYGDALES

YELLOW LAKE MEMBER:

- 1a MOSTLY PYROXENE-RICH MAFIC PHONOLITE LAVA WITH LOCAL WELL-DEVELOPED PHENOCRYSTS OF RHOMB-ANORTHOCLASE AND SOME PRIMARY ANALCITE, ABUNDANT ZEOLITE FILLINGS IN CRACKS AND AMYGDALES

- 1b PURPLE AND GREY VOLCANIC WACKE FROM EROSION OF 1a AND PINK RADIOACTIVE FELDSPATHIC TRACHYTIC ASH FLOW, SANDSTONE, AND CONGLOMERATE

- 1c CLARK CREEK PORPHYRY: A SILL-LIKE BODY RELATED TO 1a WITH LARGE FELDSPAR PHENOCRYSTS

SPRINGBROOK FORMATION

- 0a POLYMIC TIC CONGLOMERATE AND BRECCIA WITH CLASTS DERIVED MAINLY FROM PRE-TERTIARY BEDDED ROCKS

KETTLE RIVER FORMATION

- 0b MAINLY GRANITE BOULDER CONGLOMERATES, ARKOSE, VOLCANIC WACKE, AND RHYOLITE BRECCIA

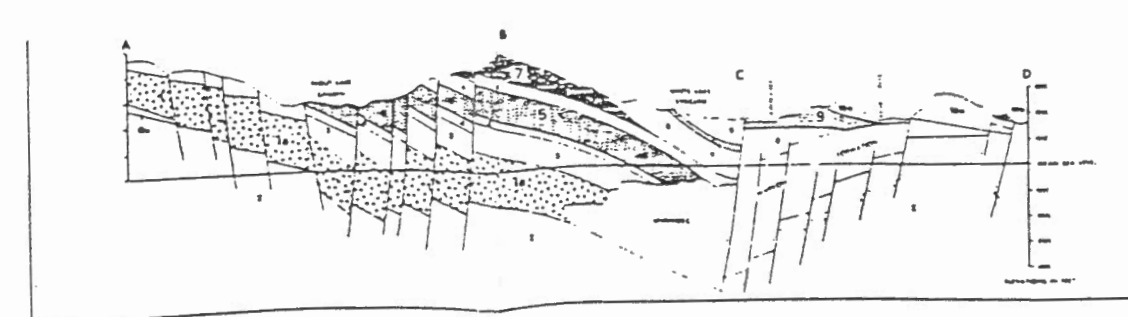
- 0c SHINGLE CREEK PORPHYRY: A COARSE SANIDINE QUARTZ PORPHYRY INTRUSION FEEDER TO THE RHYOLITE VOLCANIC ROCKS OF 0b

PRE-TERTIARY ROCKS

- Y MAINLY GRANITIC INTRUSIONS

- Z MAINLY CHERTS, GREENSTONES, SCHISTOSE ROCKS, AND MINOR INTRUSIONS

CROSS-SECTION



(from Church et al. 1983)

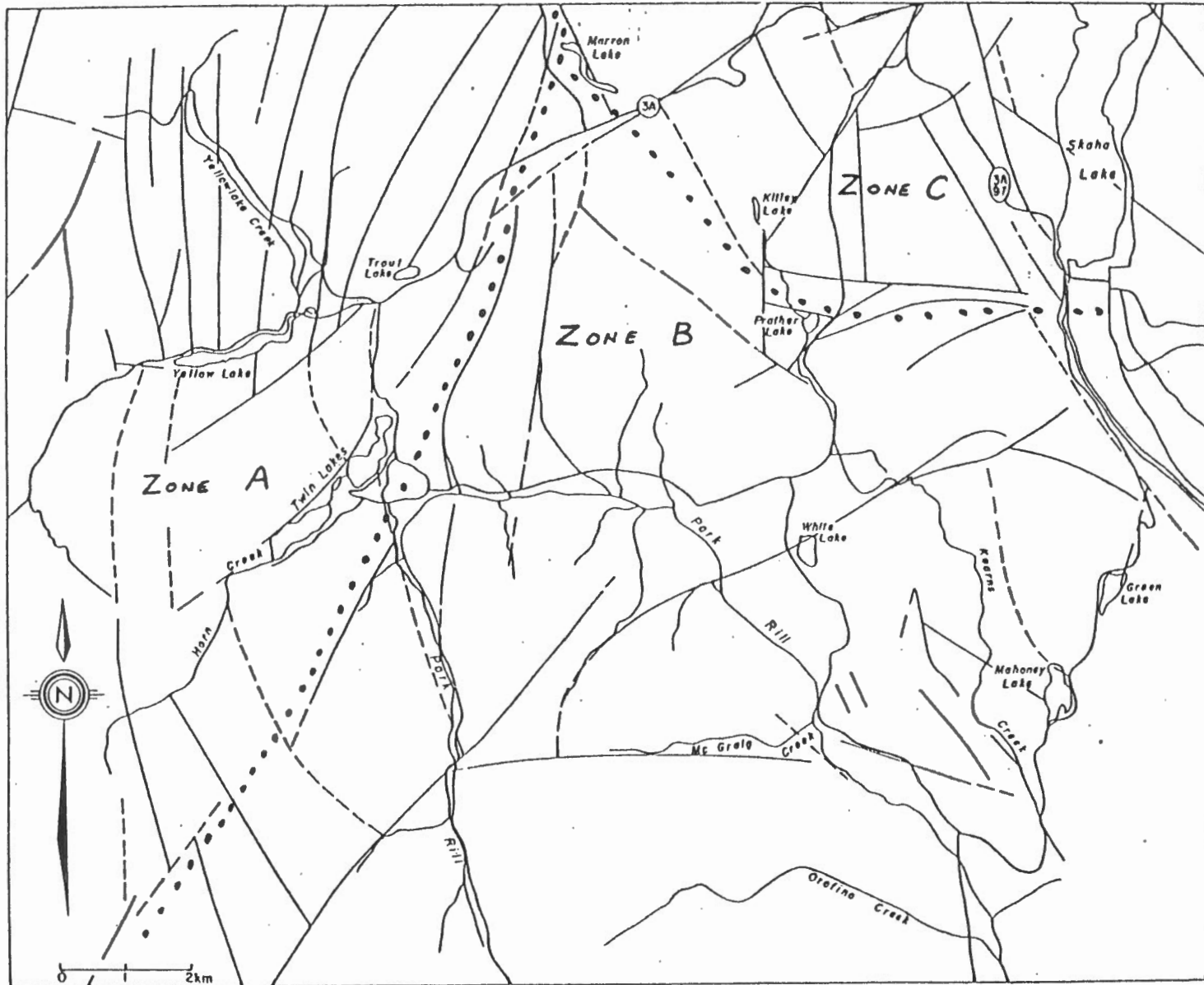


Figure 3: Distribution of known and inferred major faults in the White Lake Basin.

Skaha Formation: Slide breccia with some intercalated conglomerate and tephrite overlain by a fanglomerate.

White Lake Formation: A thick sequence of volcanic sandstones and lavas overlain by pyroclastics, volcanic breccias and sedimentary rocks containing numerous coal seams.

Marama Formation: Rhyolite and rhyodacite lava with some pyroclastic rocks and local basal conglomerate.

Marron Formation: A thick sequence of volcanic lavas ranging in composition from trachyte to andesite.

Springbrook Formation: Mainly boulder conglomerate overlying valley talus with fragments of underlying pre-Tertiary rocks.

Overburden in the area, as determined from water well records, consists of sandy clay, till, glacial fluvial sands and gravel, and talus in various combinations and thicknesses depending on location. Unconsolidated glacial deposits are restricted to valley floors and range from 15 - 60 m thick. Unsorted silts, sands and gravels cover the floors of small tributary valleys and thin talus layers mantle the upper valley slopes. Around White Lake, silt and silty clay layers are underlain by sand, gravel and rock fragments. Bedrock is exposed in a number of places, especially along cliff faces, road cuts and stream valleys.

2.2. Faulting in the Area

The complex system of faults in the White Lake Basin is shown in Figure 3. Church (1973) subdivides the area into 3 structural zones. Zone A, located in the northwest of the study area, is bounded by strong gravity faults originating in the Marron Lake area, extending northwest along the Marron Valley and wouthwest into the Similkameen Valley. The Trout Lake Graben, an area of north-south trending gravity faults, is located along the eastern margin of the zone, adjacent to the Marron Lake Fault System.

Zone B, located in the central and southern section of the study area, is bound by the aforementioned gravity fault extending into the Similkameen Valley, and a weak, easterly trending extension of the Marron Lake Fault System which passes into the central Okanagan Valley. This zone contains many gravity faults with varying trends. A northeasterly trending fault is thought to exist beneath White Lake.

Zone C, located in the northeastern section of the study area is bound by the northerly and easterly trending branches of the Marron Lake Fault System. The eastern boundary, represented by the main Okanagan Valley, is a strong northerly trending reverse fault.

2.3 Hydrostratigraphy

Potential aquifers in overburden and bedrock have been determined through examination of water well records and Church's (1973, 1979) geological descriptions.

Overburden materials from the basin generally consist of clay, sand, gravel and broken bedrock. The sands, gravels and broken bedrock have the potential to act as aquifers if hydrological conditions are favorable (i.e., good porosity, permeability, etc.). The thickest accumulations of overburden, and hence the best potential aquifers, occur on the valley floors. The clays and silty sands will act as aquitards.

Rocks of the Penticton Tertiary Outlier consist of thick volcanic sequences interlayered with clastic sedimentary and pyroclastic rocks. The thickest volcanics are feldspar porphyry lavas, lahars and pyroclastic rocks. Permeability in these rocks is probably provided by fractures associated with either fault zones or the tops of lava flows which cooled so quickly that differential contraction caused entire horizons of jointed rock (Lewis, 1984). The more felsic volcanic rocks weather quickly, producing clays that can seal permeable horizons.

Potential aquifers in sedimentary horizons of individual formations are, in order of increasing age:

Skaha Formation: Conglomerate and sandstone overlain by conglomerate constitute the main permeable horizons in this formation.

White Lake Formation: Breccias, conglomerates and sandstones occur as a thick sequence providing possible aquifers.

Marama Formation: Local basal conglomerates provide the only favorable sedimentary aquifer materials.

Marron Formation: The thin interbeds of conglomerate sandstone and soil that exist between the many volcanic flows provide possible permeable horizons.

Springbrook Formation: Thick accumulations of polymictic boulder conglomerates and breccias are potential aquifers.

Faults in the Basin are possible routes of water migration, particularly where the walls have been brecciated. Water can migrate up, down, laterally along or across faults depending on the hydrological conditions. The nature of the topography and geology in the White Lake Basin tends to create artesian pressure in the flow systems which will cause groundwater to migrate upward or along faults. The potentiometric conditions, which are influenced primarily by the elevation of the recharge area, are the most important factors controlling the direction of water flow along faults.

The cross-section drawn by Church (1979) (Fig. 2) shows the faults in the area of Yellow and Trout Lakes which form the Trout Lake Graben. These faults extend 1540 - 2150 m below the surface into the pre-Tertiary greenstones and schists. Groundwater flow in the area is probably controlled by the graben because of the number of faults and different units encountered across strike at any given depth.

In the vicinity of White Lake, the fault passing under the lake could act as a conduit to the main Okanagan Valley for waters recharging in the surrounding hills and migrating towards the lake. Unsuccessful attempts to find water at depths below White Lake suggests that the sands and gravels in the valley are drained by the fault.

Chapter 3: INTRODUCTION TO CHEMISTRY, ISOTOPES AND GEO THERMOMETRY

3.1 Water Chemistry

Dissolved inorganic constituents provide clues as to a water's influence on the rock or soil units through which it has passed.

Selected water samples were analysed chemically for Na^+ , K^+ , Ca^{2+} , Mg^{2+} , Sr^{2+} , Fe^{2+} , Mn^{2+} , HCO_3^- , Cl^- , SO_4^{2-} and SiO_2 . The anions Cl^- , HCO_3^- and SO_4^{2-} and the cations Mg^{2+} , Ca^{2+} , K^+ and Na^+ normally comprise more than 90% of the dissolved solids in water and are referred to as the 'major' ions. The concentrations of major, minor and trace inorganic constituents in groundwater is determined by the availability of the elements in the soil and rock units through which the water has passed and the interaction of these dissolved constituents with the soil and rocks units.

3.1.1 Chemistry of Water in Crystalline Rocks

Crystalline rocks of igneous origin contain considerable amounts of quartz, feldspars and micas. These minerals were originally formed at temperatures far greater than those that exist at or near the Earth's surface. "In the groundwater zone, these minerals

become thermodynamically unstable and tend to breakdown when in contact with water." (Freeze and Cherry, 1979, p. 268). This causes the water to acquire dissolved constituents.

"CO₂-charged groundwater is chemically aggressive when in contact with feldspars, micas and other silicate minerals. When CO₂-charged waters, that are low in dissolved solids, encounter silicate minerals high in cations, aluminum and silica, cations and silica are leached, leaving clay minerals such as kaolinite, illite and montmorillonite. Cations such as Na⁺, Ca²⁺, Mg²⁺ and K⁺ are released to the water. Another consequence is an increase in pH and HCO₃⁻ concentration." (Freeze and Cherry, 1979, p. 268).

In general, groundwater flow through rocks of igneous origin tends to breakdown silicate minerals, releasing ions to the water which results in alterations to the water chemistry.

3.1.2 Chemistry of Waters in Glacial Deposits

The complex mixtures of mineralogical assemblages in glacial deposits results in varied chemistries in groundwaters from these deposits.

Freeze and Cherry (1979) outline three categories of groundwaters in glacial deposits: Type I waters are slightly acidic, fresh waters with low TDS contents

(100 mg/l) in which Ca^{2+} , Na^+ and/or Mg^{2+} are the dominant cations and HCO_3^- is the dominant anion. This water is typically found in the glacial deposits of the Canadian Shield. Type II waters are slightly alkaline, fresh waters (1000 mg/l TDS) in which Ca^{2+} and Mg^{2+} are the dominant cations and HCO_3^- is the dominant anion. Type III waters are slightly alkaline, brackish waters (1000 to 10000 mg/l TDS) in which Na^+ , Ca^{2+} , Mg^{2+} , HCO_3^- and SO_4^{2-} generally occur in major concentrations.

3.1.3 Anion Evolution

Chebotarev (1955) introduced the concept of groundwater anion evolution after noting that groundwater tends to evolve chemically towards the composition of seawater: $\text{HCO}_3^- \rightarrow \text{HCO}_3^- + \text{SO}_4^{2-} \rightarrow \text{SO}_4^{2-} + \text{HCO}_3^-$
 $\text{SO}_4^{2-} + \text{Cl}^- \rightarrow \text{Cl}^- + \text{SO}_4^{2-} \rightarrow \text{Cl}^-$.

The evolution sequence occurs as water increases in age, travelling from shallow zones of active flushing through intermediate zones and on to sluggish, deeper zones. The process of evolution from stage to stage is controlled by the availability of minerals containing these ions along the flow paths, and water-rock interactions involving these ions.

3.1.4 Cation Exchange

Cation exchange is an important process that takes place between groundwater and clay minerals of colloidal size (10^{-3} - 10^{-6} mm). These clay minerals generally occur in overburden, fine-grained sedimentary rocks (shales) or as secondary alteration minerals that infill fractures or are the result of the weathering of feldspar-rich rocks. Adsorption of anions or cations is pH dependent, as a positive charge (H^+) develops on a hydrated clay mineral at a low pH, and a negative charge (OH^-) develops at a high pH.

Exchange reactions in groundwater systems involving monovalent and divalent cations such as $Na^+ - Ca^{2+}$, $Na^+ - Mg^{2+}$, $K^+ - Ca^{2+}$ and $K^+ - Mg^{2+}$ are important in determining water chemistries in clay mineral-bearing soils and rocks.

An important reaction in many groundwater systems is between Na^+ and Ca^{2+} . The exchange reaction for these two cations is



where (ad) represents an adsorbed ion. Through this reaction, Ca^{2+} -rich waters can be transformed into Na^+ -rich waters.

In general, the divalent cations will be preferentially adsorbed onto the clay minerals while the monovalent cations dominate the water chemistry.

3.1.5 Carbon dioxide in Groundwater

"Groundwater contains dissolved gases as a result of (1) exposure to the Earth's atmosphere prior to infiltration into the subsurface environment, (2) contact with soil gases during infiltration through the unsaturated zone, or (3) gas production below the water table by chemical or biochemical reactions involving the groundwater, minerals, organic matter and bacterial activity." (Freeze and Cherry, 1979, p. 95).

Carbon dioxide is one of the most important dissolved gases in groundwater. A reaction between CO₂ and water is



where (g) and (aq) denote gaseous and dissolved phases, respectively. Dissolved CO₂ in water is usually referred to as H₂CO₃ (carbonic acid). H₂CO₃(aq) is a weak acid that promotes dissolution during mineral-water interaction, thereby altering the chemistry of the waters.

$\text{H}_2\text{CO}_3(\text{aq})$ will dissociate to form HCO_3^- and CO_3^{2-} ; the relative concentrations depending upon the pH of the water. Most natural waters have a pH within the range of 6.4 to 10.3 which means that HCO_3^- will be the dominant species present. Provided that sufficient Ca^{2+} is present in the water, the introduction of HCO_3^- will result in the precipitation of calcite once calcite saturation is reached.

3.2 Geothermometry

The temperature of water in geothermal systems generally increases with depth due to the geothermal gradient within the Earth. Geothermometers represent the equilibria of temperature-dependent reactions taking place between groundwater and rocks. Groundwater may retain chemical or isotopic evidence of temperature conditions at depth provided the water has attained equilibrium with the rocks. If a geothermometer is to be accurate, not only must equilibrium conditions be reached between the ions and the water, but the water must rise relatively rapidly to the surface without alteration due to processes such as evaporation, precipitation or mixing.

3.2.1 The Na - K - Ca Geothermometer

Studies on Na^+ and K^+ species in hot water systems

suggest that alkali concentrations are controlled by equilibria with K-feldspars, plagioclase and micas (White, 1965; Ellis and Mahon, 1967). Fournier and Truesdell (1973) suggested that Ca^{2+} in high concentrations also plays a role in alkali-mineral exchange reactions, and derived an equation defining the equilibrium constant K, where:

$$\log K = \log \frac{(\text{Na}^+)}{(\text{K}^+)} + \beta \log \frac{(\text{Ca}^{2+})}{(\text{Na}^+)}$$

where $\beta = 1/3$ when $\log \frac{(\text{Ca}^{2+})}{(\text{Na}^+)}$ is negative and

$\beta = 4/3$ when $\log \frac{(\text{Ca}^{2+})}{(\text{Na}^+)}$ is positive.

The log K value obtained is then compared to measured temperatures from known geothermal fields.

Evaporation of water, causing increases in the concentrations of dissolved constituents, and subsurface mixing could change the chemistry of the sample so that equilibrium conditions are not represented and, therefore, render this geothermometer useless.

3.2.2 The Silica Geothermometer

The silica geothermometer assumes that equilibrium has been reached between water and chalcedony in low temperature systems. The temperature at the point of sampling and the concentration of silica in water can be plotted on a silica geothermometer graph and transferred to either chalcedony or quartz equilibrium curves whereby the theoretical equilibrium temperature at depth can be determined. This assumes a rapid, non-mixing ascent of the water and no loss of SiO_2 due to precipitation.

3.3 Tritium in Groundwater

Tritium, ^3H , is a radioactive isotope of hydrogen. It has an atomic mass of three, a half-life of 12.43 years and emits beta (β) radiation. Tritium occurs naturally in the atmosphere as a result of the interaction of cosmic ray-produced neutrons with nitrogen. Since 1952, tritium has been produced in large amounts as a by-product of the testing of nuclear weapons in the atmosphere.

The tritium concentration in water is measured in Tritium Units (T.U.), where 1 T.U. is equal to one tritium atom in 10^{18} hydrogen (^1H) atoms.

Natural tritium levels are usually between 5 and 20 T.U. (Brown, 1961). However, due to atmospheric nuclear

testing, tritium concentrations in precipitation reached a peak in 1962-63 at levels exceeding 100 times the natural background values. Levels have declined steadily since 1963 when the U.S.A. and U.S.S.R. placed a moratorium on atmospheric testing. However, levels increased slightly in 1967-68 when France and China did conduct some testing.

Tritium vapour is returned to the Earth's surface as precipitation. The tritium concentration in rain increases from low to high latitudes due to atmospheric circulation patterns. At a given latitude, continental precipitations are enriched in ^3H with respect to oceanic rains because of the dilution of atmospheric vapour by oceanic water vapour, which has a negligible tritium content.

Tritium is valuable as a conservative tracer in groundwater because, as it is an isotope of hydrogen, it is incorporated into the water molecule. In groundwater systems, tritium is usually employed as a guide to groundwater age. In general, if no tritium is detected, the water can be estimated to be greater than 30 years old. If some tritium is detected, the water is either younger than 1952 or is a mixture of old and young waters.

If the tritium concentration of groundwater for a specific sampling point is compared with the known pattern of tritium at the recharge area (accounting for decay), an estimate of the time period during which the

precipitation entered the groundwater system can be made.

Groundwater age is determined by the time of travel from the recharge area to the sampling point. An approximate groundwater velocity can also be inferred, provided the recharge area is well defined and the length of flow path is known.

3.4 ^2H and ^{18}O in Groundwater

Oxygen-18 and deuterium (^2H) occur naturally as part of the water molecule. Because of the variations in abundance of ^2H and ^{18}O in precipitation, they can be used as tracers in groundwater studies.

Isotope ratios of oxygen ($^{18}\text{O}/^{16}\text{O}$) and hydrogen ($^2\text{H}/^1\text{H}$) are expressed in delta units (δ) as parts per mille ($^{\circ}/\text{oo}$) differences relative to a standard. For both oxygen and hydrogen the standard used is referred to as SMOW (Standard Mean Ocean Water), where

$$\delta^{18}\text{O}_{\text{SMOW}} \quad \text{and} \quad \delta^2\text{H}_{\text{SMOW}} = 0^{\circ}/\text{oo}$$

"When water evaporates from the oceans, the water vapour produced is depleted in ^2H and ^{18}O relative to ocean water, by about $80\text{--}120^{\circ}/\text{oo}$ in ^2H and $12\text{--}15^{\circ}/\text{oo}$ in ^{18}O ." (Freeze and Cherry, 1979, p. 138.)

The process by which the isotope content of a water

changes due to condensation, evaporation, freezing and/or chemical or biological processes is known as isotope fractionation. Isotope fractionation is temperature dependent, so that isotopic composition is also temperature dependent. This makes for strong seasonal variations in isotopic compositions of precipitation for a given location.

Global ^2H and ^{18}O concentrations in precipitation surveys correlate according to the relation (Dansgaard, 1964)

$$\delta ^2\text{H}^{\circ}/\text{oo} = 8 \delta ^{18}\text{O}^{\circ}/\text{oo} + 10$$

which is known as the global meteoric water line. This exact relationship changes slightly according to location and precipitation pattern. There are several basic principles associated with isotopic concentrations in precipitation (Fontes, 1979):

(1) Continental precipitation is depleted in ^2H and ^{18}O as compared to marine and coastal precipitations.

(2) Precipitation falling under cooler climatic conditions contain lower concentrations of the heavy isotopes as compared to warmer conditions.

(3) Winter precipitation is depleted in ^2H and ^{18}O with respect to summer precipitation.

(4) There is a negative linear correlation of heavy isotope content of precipitation with altitude.

(5) When water is evaporated under natural conditions, the remaining water becomes enriched in ^2H and ^{18}O .

Chapter 4: DISCUSSION AND INTERPRETATION

4.1 Field Program

Visits were made to the White Lake Basin to collect water samples from selected surface waters, springs, domestic wells and diamond drill holes during mid-May and early September, 1983. Many of the sampling sites were the same as those sampled by Michel and Fritz in October, 1981. Periodic sampling at the same sites provides data for investigating seasonal or yearly variations in groundwater properties.

A total of fifty-nine samples were collected, ten from diamond drill holes, nineteen from domestic wells, nine from springs and twenty-one from surface water bodies, during the two sampling periods. Diamond drill holes were sampled at various depths to determine variations in groundwaters entering the holes at different levels. A stainless steel sampler, attached to 615 m of stainless steel wire, was lowered down the wells to the required sampling depths using a portable winch.

Conductivity, salinity, pH and temperature were measured in the field, while samples were collected for isotope and chemical analysis in the laboratory. Figure 4 shows the location of sampling sites and Table 1 contains a listing of sampling locations and field

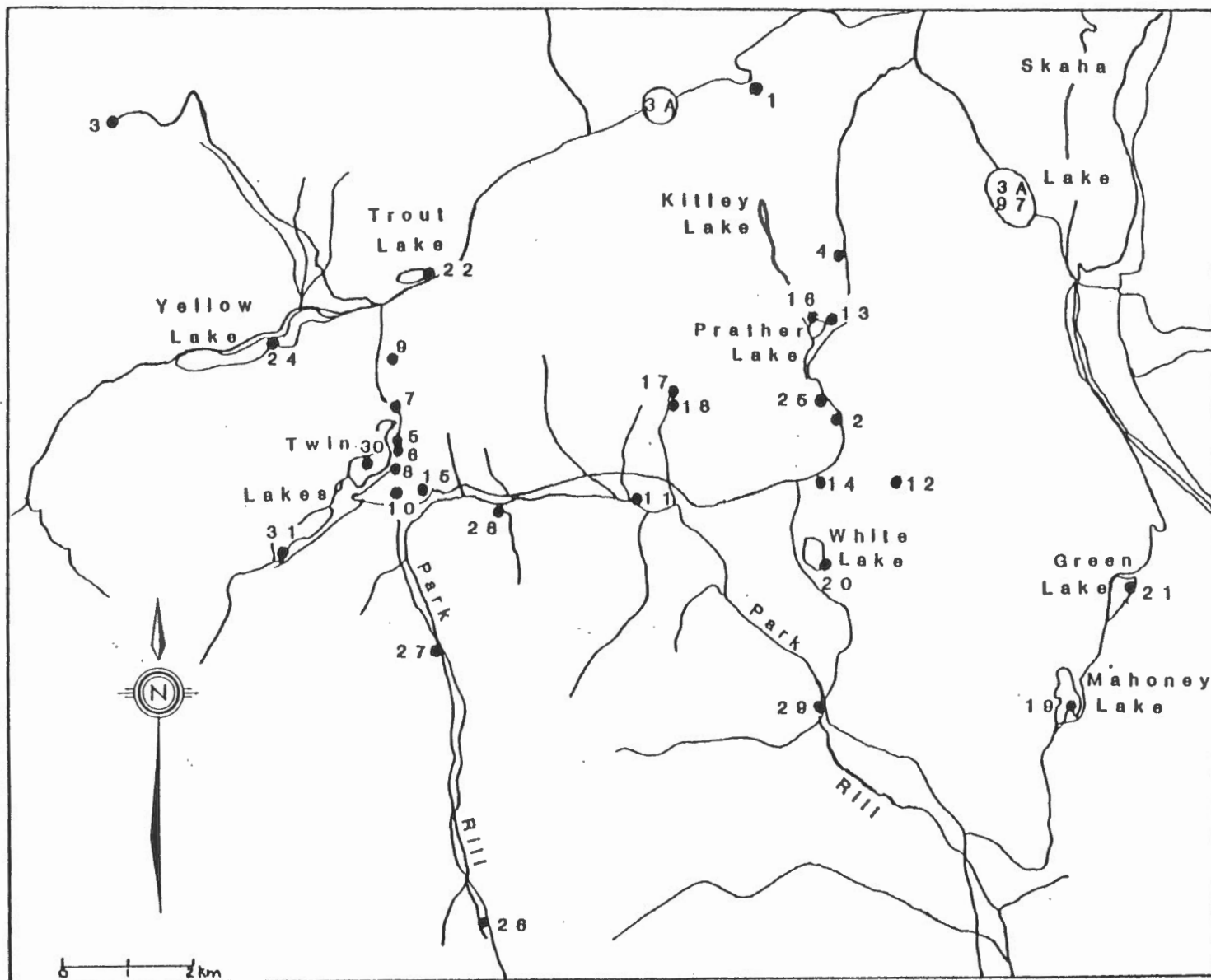


Figure 4: Location of sampling points.

TABLE 1: FIELD MEASUREMENTS

Sample No.	Sample Name	Temperature (°C)	Conductivity (Microsiemens /cm)	Salinity (°/oo)	pH
<u>DIAMOND DRILL HOLES</u>					
1a	78-4 (surface)	13.5	2020	1.5	7.70
1b	78-4 (154 m)	18.0	2000	1.0	7.55
1c	78-4 (385 m)	18.2	2130	1.2	7.75
1d	78-4 (430 m)	18.9	2190	1.1	7.95
2a	P-Well (surface)	10.3	1550	1.1	7.95
2b	P-Well (41 m)	14.0	1700	1.3	8.00
2c	P-Well (492 m)	-	-	-	-
3	78-3 (18 m)(1981)	11.0	-	-	-
<u>DOMESTIC WELLS IN BEDROCK</u>					
4	Mr. Bork	12.8	785	0.2	8.25
5	Mr. Nigilser(178 m)	17.5	279	0.1	9.30
6	Mr. Nigilser(173 m)	14.9	1200	0.8	7.90
<u>DOMESTIC WELLS IN OVERBURDEN</u>					
7	Mr. Bohn	15.0	550	0.0	7.20
8	Mr. McWhinnie	10.0	483	-	7.55
9	Twin Lake Golf Club	17.5	620	0.4	7.40
10	Twin Lake Ranch	12.8	443	-	7.45
11	White Lake Ranch#2	15.0	383	0.1	6.65
12	Observatory	19.0	600	0.1	-
13	St.Andrews Golf Club	17.0	740	0.5	6.65

TABLE 1: FIELD MEASUREMENTS

Sample No.	Sample Name	Temperature (°C)	Conductivity (Microsiemens /cm)	Salinity (°/oo)	pH
<u>SPRINGS</u>					
14	Observatory	10.9	680	0.5	7.23
15	Twin Lake Ranch	17.0	990	0.5	7.05
16	St.Andrews Golf Club	9.5	322	0.0	7.25
17	White Lake Ranch#1	15.2	218	0.0	-
18	White Lake Ranch#2	10.8	210	-	8.02
<u>SURFACE WATERS</u>					
19	Mahoney Lake	20.7	8500	5.2	8.35
20a	White Lake(Aug/83)	23.0	4150	2.5	9.00
20b	White Lake(May/83)	19.5	2100	1.1	9.00
21	Green Lake	19.8	2300	2.2	8.95
22	Trout Lake	19.0	600	0.3	7.35
23	Farleigh Lake	18.8	285	0.0	8.50
24	Yellow Lake	19.0	370	0.4	8.85
25	St.Andrews Golf Club				
	Stream	17.0	590	0.3	6.40
26	Park Rill #1	9.0	325	0.1	-
27	Park Rill #2	12.0	510	0.2	7.80
28	Park Rill #3	14.0	600	0.2	7.70
29	Park Rill #4	13.0	610	0.3	7.05
30	Twin Lake #1	20.0	402	0.0	8.30
31	Twin Lake #2	19.5	254	0.0	7.85

measurements.

The diamond drill holes sampled, 78-3, 78-4 and P-Well are all greater than 300 m in depth. Boreholes 78-3 and 78-4 were drilled as part of a uranium exploration program while P-Well was drilled as part of the study by Jessop and Judge (1971) of heat flow across Canada.

The domestic wells sampled have been subdivided according to whether they were completed in bedrock or overburden. The springs sampled were usually located near the valley floor. They were conspicuous by the wet ground and amount of vegetation surrounding them in an otherwise very dry and barren area.

All samples for isotopic and chemical analysis were sent to the Environmental Isotope Laboratory at the University of Waterloo.

4.2 Discussion of Results

4.2.1 Tritium

In this study, tritium concentrations in precipitation from Ottawa, Canada, have been assumed to be similar, or slightly higher than those from the White Lake Basin. This assumption is based on a map showing tritium distributions in precipitations from North

America (Michel, 1977, p. 33). Measured tritium concentrations can be compared to known yearly averages for the Ottawa area (Figure 5) and a rough approximation of age can be made. Precipitation samples from the Penticton Airport currently contain tritium concentrations ranging from 26 to 45 T.U. Tritium data are listed in Table 2.

For most samples collected from diamond drill holes, tritium was not detected, indicating that the waters are greater than thirty years old. Tritium was detected in samples from two holes. Water from 78-3 (surface) and 78-4 (154 m) contains 21 and 9 T.U., respectively. These measurable values could be the result of downhole mixing with younger waters. Lewis (1984) shows a marked decrease in water temperature in 78-4 at a depth of approximately 350 m. This could be the result of younger, colder waters entering the hole through one of the many faults in the basin. This increase in tritium does not affect the water sampled at 385 m and, hence, tritium was not detected in water from this depth. The water from hole 78-3, sampled in 1981, could also be the result of mixing between old, deep waters and younger, shallower, fault-related waters.

It should be emphasized at this point that the tritium analysis has an error factor of ± 10 T.U. and that the low values from 78-3 and 78-4 may be very low or zero, indicating pre-1952 water.

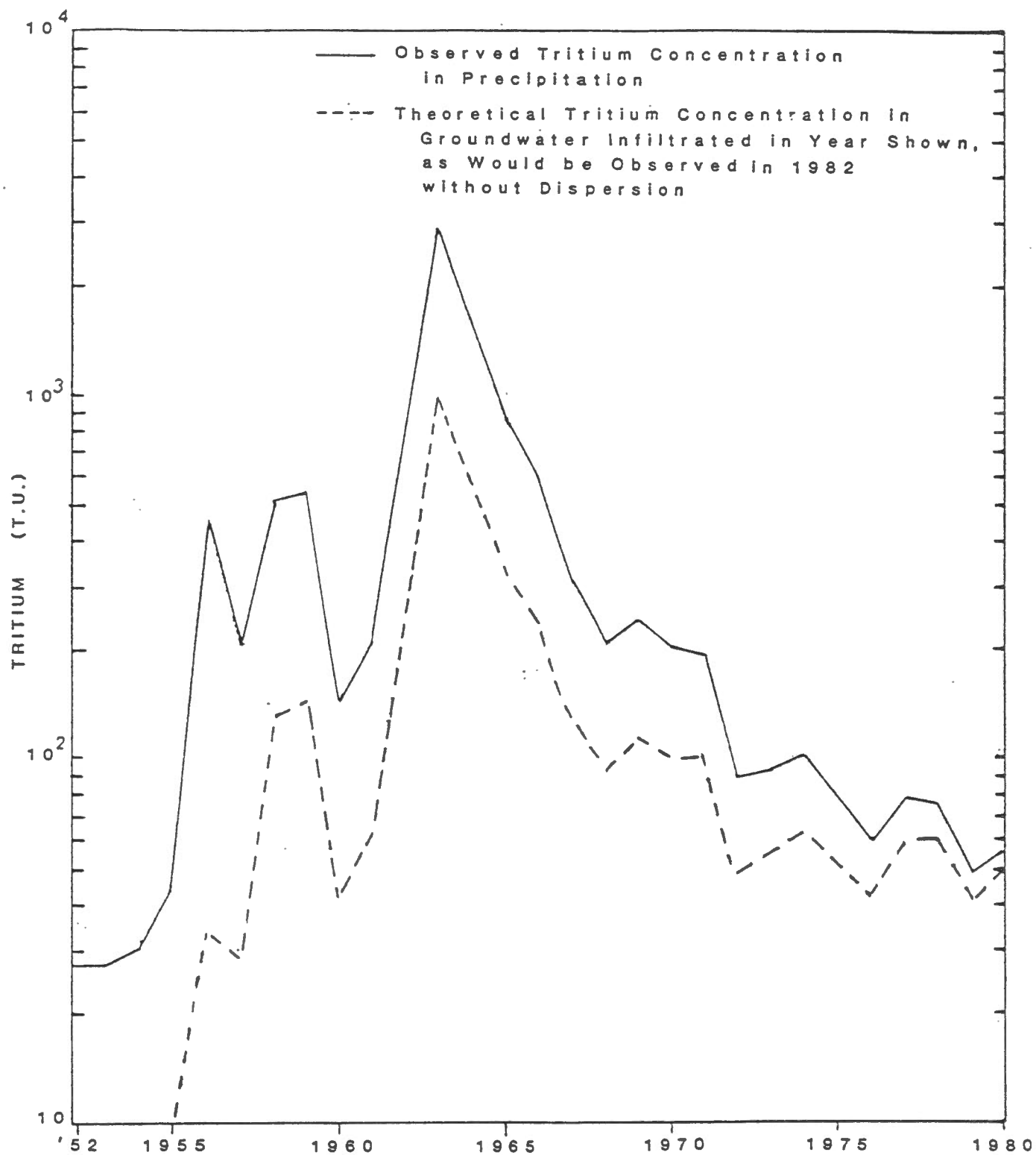


Figure 5: Mean annual tritium concentrations in precipitation and groundwaters at Ottawa, 1952-1980. (Modified after Kubasiewicz, 1983).

TABLE 2: ISOTOPE DATA OF WATER SAMPLES

Sample No.	Sample Name	$^{18}\text{O}/\text{OO}(\text{SMOW})^1$	$^2\text{H}/\text{OO}(\text{SMOW})^2$	$^3\text{H}(\text{T.U.})^3$
<u>DIAMOND DRILL HOLES</u>				
1a	78-4 (surface)	-18.2	-138	n.d.
1b	78-4 (154 m)	-18.0	-140	+9
1c	78-4 (385 m)	-18.2	-135	-
1d	78-4 (430 m)	-18.2	-134	n.d.
2a	P-Well (surface)	-18.1	-139	n.d.
2b	P-Well (41 m)	-18.1	-137	n.d.
2c	P-Well (492 m)	-17.9	-132	n.d.
3	78-3 (18 m)(1981)	-17.0	-133	+21
<u>DOMESTIC WELLS IN BEDROCK</u>				
4	Mr. Bork	-16.0	-133	n.d.
5	Mr. Nigilser(178 m)	-18.4	-149	-
6	Mr. Nigilser(173 m)	-16.2	-133	+77
<u>DOMESTIC WELLS IN OVERBURDEN</u>				
7a	Mr. Bohn (Sept/83)	- 7.8	- 97	+56
7b	Mr. Bohn (May/83)	- 7.7	- 96	-
8	Mr. McWhinnie	-13.7	-119	n.d.
9	Twin Lake Golf Club	- 9.7	-103	n.d.
10	Twin Lake Ranch	-14.0	-126	+48
11	White Lake Ranch#2	-15.6	-126	+40
12	Observatory	-15.3	-130	+50

TABLE 2: ISOTOPE DATA OF WATER SAMPLES

Sample No.	Sample Name	$^{18}\text{O}^{\circ}/\text{OO}(\text{SMOW})^1$	$^2\text{H}^{\circ}/\text{OO}(\text{SMOW})^2$	$^3\text{H}(\text{T.U.})^3$
<u>SPRINGS</u>				
14a	Observatory(Aug/83)	-16.2	-133	+26
14b	Observatory(May/83)	-15.9	-131	-
15	Twin Lake Ranch	-12.5	-116	+33
16	St. Andrews Golf Club (Aug/83)	-16.2	-129	+24
16b	St. Andrews Golf Club (May/83)	-16.0	-129	-
17	White Lake Ranch#1	-16.5	-132	+17
<u>SURFACE WATERS</u>				
19	Mahoney Lake	- 6.5	- 87	-
20a	White Lake(Aug/83)	- 5.3	- 81	+30
20b	White Lake(May/83)	-13.4	-118	-
21	Green Lake	- 6.0	- 83	-
22	Trout Lake	- 8.4	- 94	-
26	Park Rill #1	-16.1	-126	+67
28	Park Rill #3	-15.9	-132	-
30a	Twin Lake #1 (Aug/83)	- 8.3	- 98	-
30b	Twin Lake #1 (May/83)	- 9.1	- 98	-
31	Twin Lake #2	-11.6	-106	-

TABLE 2: ISOTOPE DATA OF WATER SAMPLES

Sample No.	Sample Name	$^{18}\text{O}/\text{oo}(\text{SMOW})^1$	$^2\text{H}/\text{oo}(\text{SMOW})^2$	$^3\text{H}(\text{T.U.})^3$
------------	-------------	--	---------------------------------------	-----------------------------

PRECIPITATION

June, 1983		-12.7	-100	+26
July 1-8, 1983		-13.4	-109	+29
July 8-25, 1983		- 9.0	- 84	+45
August, 1983		- 7.1	- 63	+28
September, 1983				
October, 1983				
November, 1983				
December, 1983				

1) Analytical precision is $\pm 0.2^\circ/\text{oo}$

2) Analytical precision is $\pm 2^\circ/\text{oo}$

3) Analytical precision is ± 10 T.U.

n.d. = not detected

The two bedrock wells owned by Mr. Nigilser both yielded water containing tritium. Water from the 178 m and 173 m deep wells contain 12 and 77 T.U., respectively. The water from the 178 m well was probably recharged during the mid to late 1950's, whereas water from the 173 m well was probably recharged during the early 1970's. It is possible that these two wells are intercepting water from different flow systems despite their close proximity to each other.

Springs in the area are discharging relatively young waters, with tritium contents ranging from 17 to 33 T.U. These waters are probably less than 5 years in age.

White Lake and Park Rill #1 were the only surface waters tested for tritium, and they contain 30 and 67 T.U., respectively. The White Lake sample is probably indicative of groundwater less than 5 years old. The Park Rill #1 sample was taken close to the source of the stream, which originates from groundwater seepage. This seepage is probably representative of a shallow flow system that was recharged in the early to mid 1970's, which explains the higher tritium value.

Wells completed in overburden have consistent tritium values, ranging from 40 to 56 T.U. This suggests that waters from these wells are part of relatively short, shallow flow systems that recharged around the mid 1970's. The exceptions to this are the

wells owned by Mr. McWhinnie and the Twin Lakes Golf Course, where tritium was not detected. These waters are greater than 30 years old and must be part of a longer, slow moving flow system within the overburden.

It is possible to distinguish between waters from deep water bodies and the shallow and surface water bodies on the basis of tritium contents. Tritium is not detected in most diamond drill holes, whereas there is detectable tritium in most wells, springs and surface water samples. It is difficult to distinguish between the shallow and surface water bodies on the basis of tritium, as their concentrations are very close to one another. However, it is assumed that subsurface water is older than surface water, in which case their possible range of ages can be reduced and roughly separated.

4.2.2 ^2H and ^{18}O Isotopes

The results of the isotopic analyses of waters from the White Lake Basin are presented in Table 2 and Figure 6. Data from the initial report of Michel and Fritz (1981) have been used to compliment the data from the 1983 field program.

The global meteoric water line, defined by the relation (Dansgaard, 1964) $\delta^2\text{H}^{\circ}/\text{oo} = 8 \delta^{18}\text{O}^{\circ}/\text{oo} + 10$, serves as a reference for the isotopic composition of

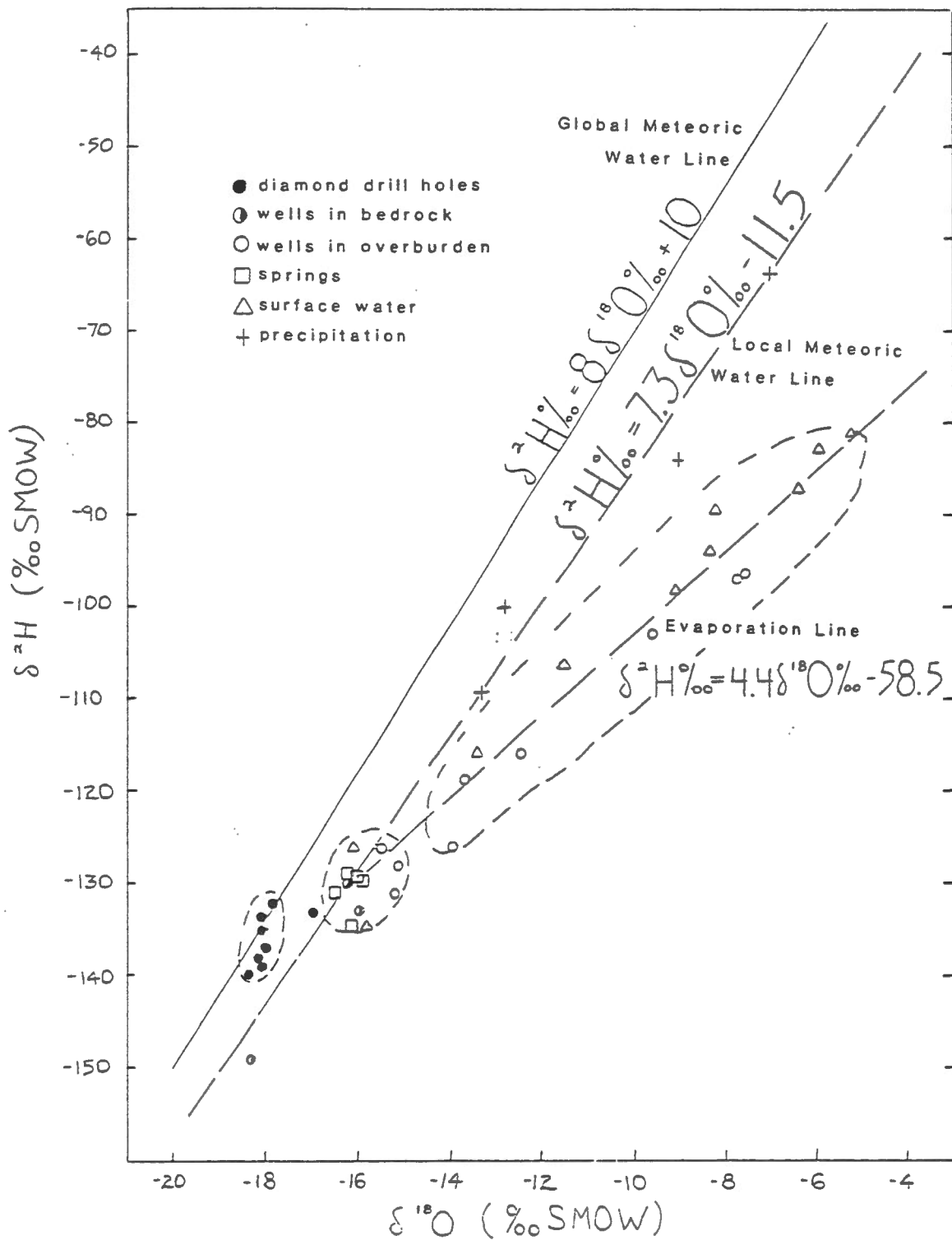


Figure 6: Oxygen and hydrogen isotopic compositions of waters of the White Lake Basin, B.C.

the waters from the basin. The local meteoric water line (or precipitation line) was determined from the isotopic compositions of various precipitation samples collected at the Penticton Airport, during the period of June to December, 1983. ^2H and ^{18}O appear to be locally associated by the relation $\delta^{2\text{H}}\text{‰} = 7.3 \delta^{18\text{O}}\text{‰} - 11.5$.

Unevaporated, shallow groundwaters that have been recharged within the White Lake Basin have an average isotopic composition of approximately (-16.2, -130). This value plots on the local meteoric water line. The water in this group, outlined in Figure 6, consists of several springs, freshwater streams and shallow domestic wells in overburden. It probably represents precipitation that has infiltrated the shallow, unconsolidated aquifers immediately or very shortly after falling. It has undergone little or no evaporation, which would have resulted in a change in isotopic composition.

Water that has not entered the subsurface immediately, has undergone some degree of evaporation. This water could have ponded on the surface prior to infiltration or become part of a lake or stream. The arid climate of the Okanagan Valley facilitates the evaporation of surface waters.

Evaporated water in the White Lake Basin is enriched in ^2H and ^{18}O and plots below the local

precipitation line in Figure 6. The evaporation line, as determined from several surface waters and shallow domestic wells, is defined by the relation $\delta^{2H^{\circ}/\text{oo}} = 4.4 \delta^{18O^{\circ}/\text{oo}} - 58.5$.

The waters that have undergone the most evaporation are White Lake, Green Lake and Mahoney Lake. There is a lack of good surface drainage in these lakes, indicating that the bulk of the water loss is by evaporation. Waters from the Twin Lakes are also slightly evaporated, although not to the degree of the aforementioned lakes.

Several shallow wells are also located on the evaporation line. The water from these wells was probably evaporated to some degree before being recharged. The well of Mr. Bohn, located just north of Twin Lake #1 in a kame deposit, contains evaporated water. This water is thought to have come directly from Twin Lake #1 because of the similarity in isotope values. There has been some component of mixing of evaporated and unevaporated waters. The Twin Lakes Golf Course well probably receives some of its water from the Twin Lakes themselves. However, water from Twin Lake #1 is considerably more evaporated than water from the golf course, indicating that there has been some mixing with less evaporated water. It is likely that the golf course receives some of its water from the steeply rising hill to the east and as direct recharge into the sand and gravel deposits located between the Twin Lakes and the golf course.

Water from the Twin Lake Ranch spring plots on the evaporation line. At the time of sampling, the field in which the spring is located was being irrigated. The water, sprayed as a fine mist, would have undergone some evaporation during its travel through the air and during the interval before it was able to infiltrate into the subsurface. The exact source of the spring was not located, and a sample was taken as close to the source as possible, in a wet, marshy area. It is likely that the sample taken contained some of the evaporated irrigation water.

The diamond drill holes contain waters that are isotopically distinct from any other waters in the basin. In Figure 6 these waters are shown to fall above the local line and very close to the global meteoric water line. These waters are depleted in heavy isotopes when compared to the shallow and surface waters.

There are two possible explanations for these isotopically distinct waters. The first is that they may have fallen as precipitation at a higher altitude in a different area. If this were the case, the climate of the area would have to be sufficiently different from that of the White Lake Basin, otherwise the water would still plot on the local meteoric water line despite the difference in altitude. For example, precipitation from Victoria (coastal) and Mount Meager (2,000 to 3,000 m) plot on the same meteoric water line despite the large differences in elevation because the precipitation at

both localities is from the same vapour mass (Clark et al. 1980).

Alternatively, the waters may have been recharged under different climatic conditions than currently exist. A cooler, wetter climate could explain the lower ^2H and ^{18}O contents. This suggests that the waters are quite old, as the last major cooler climatic period ended nearly 10,000 years ago.

Water from the two wells owned by Mr. Nigilser are isotopically different, suggesting that the waters are possibly of different origin. The 178 m well has $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of (-18.4, -149), respectively, and could be an example of water recharged at a higher altitude in the local area. The 173 m well has $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of (-16.2, -130), which is representative of the more shallow, unevaporated waters that have been discussed previously.

4.2.3 Water Chemistry

The chemistry of the groundwaters from the White Lake Basin is dependent on many factors. The original composition of the water when it fell as precipitation is the original factor. Others include length and depth of the flow system, the mineralogical composition of the reservoir material, the age of the water and whether or not evaporation has taken place.

The results of the chemical analyses are presented in Table 3 and on a Piper diagram in Figure 7. Waters from the basin with low TDS contents are of the Ca-HCO₃ or Ca-Mg-HCO₃ type as outlined in Figure 7. Most waters originate as Ca-Mg-HCO₃ type waters because of the incorporation of soil CO₂ and the dissolution of calcite. Waters in this group are from springs and wells in overburden and are similar to type II glacial waters. They are representative of short, shallow flow systems with relatively low dissolved solids contents. The amount of dissolved solids increases with increasing length of the flow system and the water chemistry plots towards the right of the Piper diagram as evolution proceeds. Concentrations of Na⁺, SO₄²⁻ and Cl⁻ increase while Ca²⁺, Mg²⁺ and HCO₃⁻ decrease. This is typical of the normal evolution of unevaporated groundwaters.

Mr. Nigilser's 173 m well (6 on the Piper diagram), which is completed in bedrock, has been included as part of this group. It lies on the path of chemical evolution from the original water, but has a higher TDS content than the original Ca-Mg-HCO₃ waters. This reflects the longer length of time that it has been in contact with the rocks.

The surface waters outlined in Figure 7 represent samples from Green and Mahoney Lakes. They are highly evaporated, as evidenced by their high TDS contents and enrichment in heavy isotopes. They are thought to follow an evolutionary path similar to that previously

TABLE 3: CHEMICAL COMPOSITION OF SELECTED WATER SAMPLES

Sample No.	Sample Name	Na ⁺	K ⁺	Ca ²⁺	Mg ²⁺	Sr ²⁺	Fe ²⁺	Mn ²⁺	HCO ₃	Cl	F	SO ₄	SiO ₂
<u>DIAMOND DRILL HOLES</u>													
1a	78-4 (surface)	500	1.59	36.4	0.56	13.7	7.70	0.05	233	501	6.30	319	7.5
1b	78-4 (154 m)	500	1.75	34.5	0.66	13.1	0.54	0.05	266	469	4.88	292	12.5
1c	78-4 (385 m)	490	2.25	34.1	0.70	12.9	0.97	0.05	167	513	4.35	305	14.3
2a	P-Well (surface)	520	2.33	16.3	1.16	2.37	4.70	0.06	1199	161	4.92	9.29	7.5
2b	P-Well (41 m)	540	3.33	17.5	1.38	3.05	5.80	0.08	1272	172	3.38	8.39	10
2c	P-Well (492 m)	180	2.70	74	0.78	-	71.4	0.29	42.6	286	4.52	111	5
3	78-3 (18 m)	-	-	-	-	1.64	-	-	-	1.23	0.64	-	-
<u>DOMESTIC WELLS IN BEDROCK</u>													
4	Mr. Bork	245	1.60	2.54	1.54	0.82	5.08	0.05	665	9.18	4.65	31.4	7.3
5	Mr. Nigilser(178 m)	69.0	0.12	2.24	0.08	0.15	0.07	0.05	122	22.6	6.29	21.1	13
6	Mr. Nigilser(173 m)	160	1.83	103	44.1	5.65	15.7	0.21	4.20	140	0.29	47.3	31
<u>DOMESTIC WELLS IN OVERBURDEN</u>													
7	Mr. Bohn	55.5	2.43	65.0	25.6	2.31	2.21	0.20	403	4.01	0.35	123	11
8	Mr. McWhinnie	52.5	2.02	55.3	28.2	2.01	1.55	0.76	381	2.37	0.72	71.9	12

TABLE 3: CHEMICAL COMPOSITION OF SELECTED WATER SAMPLES

Sample No.	Sample Name	Na ⁺	K ⁺	Ca ²⁺	Mg ²⁺	Sr ²⁺	Fe ²⁺	Mn ²⁺	HCO ₃	Cl	F	SO ₄	SiO ₂
9	Twin Lake Golf Club	45.0	2.02	78.0	37.7	2.52	1.40	0.65	381	2.69	0.4	71.9	12
11	White Lake Ranch#2	22.5	0.38	49.0	14.1	0.41	0.09	0.08	250	1.13	0.5	22.5	-
<u>SPRINGS</u>													
14	Observatory	125	2.13	49.6	12.5	3.91	0.16	0.05	494	24.9	1.08	30.2	-
16	St. Andrews G.C.	21.5	0.28	57.0	15.6	0.98	0.05	0.05	256	1.96	0.6	32.1	-
<u>SURFACE WATERS</u>													
18	White Lake Ranch Spring#2	23.5	0.30	43.0	7.25	0.26	0.92	0.05	202	0.60	0.05	13.5	-
19a	Mahoney Lake	1550	198	14.9	508	0.28	0.07	0.05	710	321	1.06	1010	-
19b	Mahoney Lake(1981)	4900	730	28.5	2225	0.44	0.17	0.07	1940	1121	n.d.	19,440	-
20a	White Lake (Aug.)	1000	83.0	17.9	23.0	2.82	1.10	0.05	1605	168	4.03	962	-
20b	White Lake (May)	520	40.4	21.0	20.7	-	0.98	0.05	820	79	2.65	432	-
21	Green Lake	228	37.0	6.08	280	0.05	0.05	0.05	1224	23.9	0.6	701	-

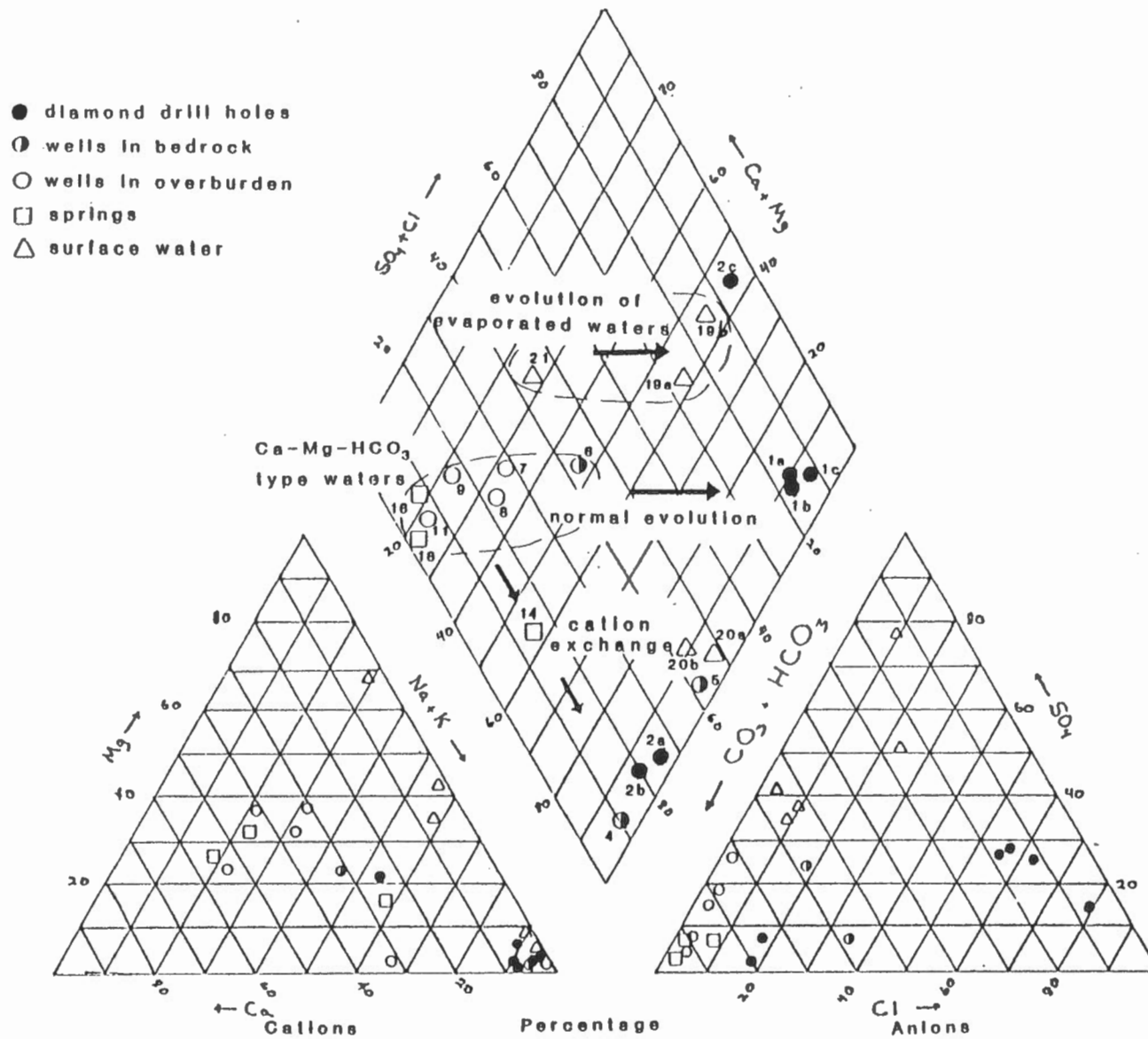


Figure 7: Major ion chemistry of waters of the White Lake Basin, B.C.

described. Minor differences in the chemical composition of the starting waters, possibly due to the different locations of the lakes in the study area, would account for the different positioning of the path on the Piper diagram. These waters are evolving, not because of their contact with rocks, but because of the large amount of evaporation taking place.

The water from Observatory Spring (14) appears to have evolved from the 'original' waters by the process of cation exchange. It probably started out as water similar in composition to that of St. Andrews Golf Club spring (16) and White Lake Ranch spring #2 (18) and, upon migration through unconsolidated sediments and underlying sedimentary rocks, it has gained Na^+ and K^+ and lost Ca^{2+} and Mg^{2+} . The TDS contents of these three springs are similar, as are the anion chemistries, but the cations have been altered.

Mr. Bork's well (4) contains Na-K- HCO_3 type water which is low in Ca^{2+} and Mg^{2+} . This well is completed in the Kitley Lake Member of the Marron Formation. This member consists of trachyandesitic lava with glomerophenocrystic clots of feldspar (Church, 1979). The cation chemistry, therefore, probably reflects the feldspar composition of the lavas. However, the chemistry could also be due to weathering of the feldspars and feldspar-rich lavas to produce clay minerals necessary for cation exchange. This cation exchange could be a further evolutionary stage of waters similar to those

from Observatory Spring (14). The TDS content of Mr. Bork's water is slightly larger than the Observatory Spring, suggesting that the well water has been in the subsurface for a slightly longer period of time.

Water from White Lake (20a and 20b) represents highly evaporated conditions, however, it does not plot with the other evaporated surface waters on Figure 7. It contains high Na^+ , K^+ and HCO_3^- and low Ca^{2+} , Mg^{2+} and Cl^- concentrations. Its chemical composition could be the result of evolution by evaporation of discharging groundwaters, which have undergone some cation exchange, similar to those from Observatory Spring (14).

Waters from P-Well were sampled at surface (2a), 41 m (2b) and 492 m (2c). The chemistry of the waters from the surface and 41 m is quite different from that of 492 m. The shallow water is a Na - HCO_3 type and the deep water is a Na - Ca - Cl type.

Despite the fact that many faults intersect P-Well at depth (Church, 1973), it is speculated that no water enters the hole through these faults because of the artesian nature of the well. If water did enter the hole at depth, it would have to be under a greater hydraulic pressure than the artesian pressure in the well at that point. It is thought, therefore, that the water from 492 m must evolve into the water from the surface and 41 m.

As water from depth flows up the hole it passes through the thick sedimentary sequence of the White Lake

Formation. There are several coal seams within these sandstones, mudstones, conglomerates and pyroclastics. The presence of bacteria in these coal seams could generate CH_4 and CO_2 , which changes to HCO_3^- because of pH conditions. This addition of HCO_3^- at depth would explain the increased concentrations in the shallower waters. The consumption and reduction of SO_4^{2-} by bacteria in the coal would also explain the decrease in SO_4^{2-} from deep to shallow waters. As the water flowed through the sedimentary rocks, cation exchange may have occurred, increasing the Na^+ and decreasing the Ca^{2+} concentrations. The water would then have roughly the same chemical composition as is seen at the shallow depths.

Waters from diamond drill hole 78-4 (1a, 1b, 1c) do not appear to be related to those from the P-Well. 78-4 waters plot to the extreme right of Figure 7, and appear to be the most highly evolved waters among those following the 'normal' evolutionary path. Their TDS contents are the highest ($\sim 2100 \mu\text{S}$) in this group. The anions in these waters as well as the cations have evolved, in contrast to the shallow water from P-Well, where only the cation chemistry has evolved from the original Ca - Mg - HCO_3 waters due to cation exchange.

The water from Mr. Nigilser's 178 m well (5) is very different chemically from his 173 m well (6) despite their close proximity. The 178 m well plots close to White Lake on Figure 7 and is a Na - HCO_3

type water. It is a very fresh water having a TDS content of only 279 μ S. The chemical history of this water is difficult to interpret. However, it is thought that the water has intersected a fault at depth. This fault, acting as a groundwater conduit giving good flow, explains the large difference in the suggested pumping rates of the two wells (173 m - 500 gpd, 178 m - 5500 gpd). The water may have started off with a composition similar to the 'original' waters, explaining the higher HCO_3^- content, and intersected the fault, where cation exchange has resulted in the low Ca^{2+} and Mg^{2+} and high Na^+ concentrations.

4.2.4 Geothermometry

The Na - K - Ca and SiO_2 geothermometers were applied to the deep waters of the White Lake Basin. The SiO_2 geothermometer predicts equilibrium temperatures in the 0 - 20°C range, while the Na-K-Ca geothermometer yields unrealistic temperatures. This suggests that the waters have not chemically equilibrated with the rocks (SiO_2) at depth or that the waters have not seen significantly higher temperatures than measured at the sampling points.

It is likely that the Na-K-Ca geothermometer is not applicable to the waters. Evaporation of the waters, causing increased concentrations of Na^+ , K^+ and Ca^{2+} ,

cation exchange, or introduction of different waters through faults or fractures would render the Na⁺ - K⁺ - Ca geothermometer useless. It is also possible that equilibrium has not yet been reached between SiO₂(aq) and the wall rocks, which would make this geothermometer useless.

4.3 Interpretations

On the basis of the chemistry, oxygen - 18, deuterium and tritium evidence presented in this study, three water systems can be delineated in the White Lake Basin. They are surface water bodies, such as lakes and streams, shallow flow systems within the overburden and shallow bedrock aquifers, and deeper, old, chemically and isotopically distinct waters.

Evaporation has taken place among some surface and shallow waters to varying degrees, as evidenced by the isotopic and chemical data. Evaporated water is typically enriched in heavy isotopes and falls below the local meteoric water line. Chemically, evaporated waters contain greater concentrations of dissolved solids than unevaporated waters.

Typically more evaporation has taken place in surface water bodies, especially Green, White and Mahoney Lakes. The lakes become highly evaporated because of the low precipitation, dry, hot summers and poor surface water flow in the area. They are generally

isolated, with little or no mixing caused by rivers or streams feeding or draining them. There is some evaporation of shallow groundwater, which has occurred prior to infiltration.

It is thought that intermediate and local flow systems existing within the shallow subsurface are influenced by topography and hydrogeological conditions (hydraulic gradients, permeability of aquifer materials, etc.). Streams, lakes and springs are the most likely discharge points for these flow systems.

The waters from diamond drill holes are distinct from the more shallow waters. They differ both chemically and isotopically.

In most diamond drill holes tritium is not detected, indicating that the waters are greater than 30 years old. This is in contrast to many of the springs and domestic wells, in which tritium was detected, suggesting the presence of at least some post-1952 water.

Chemically, the deep waters typically contain larger TDS contents, specifically Na^+ and Cl^- , than the shallow waters, with the possible exception of the evaporated lakes. This suggests that the deep waters have evolved more fully and are much older than the shallow waters.

Isotopically, the deep waters are depleted in ^2H and ^{18}O . They plot above the local meteoric water line,

suggesting that they might have been recharged under different climatic conditions than currently exist.

Individually, the waters from the diamond drill holes do not appear to be related to one another. The chemistry of the waters appear to be reflecting water-rock interactions in the vicinities of the holes. This would suggest that either there is no regional flow system in the basin or that the flow is extremely slow and any evidence of water from one hole has not yet appeared in the next hole.

On the basis of geothermal gradients, Lewis (1984) suggests the presence of a regional flow system, recharging in White Lake, and flowing up dip towards hole 78-3. None of the boreholes available for sampling penetrated to the lowest units in the Basin and, therefore, no direct evidence to support or reject Lewis' hypothesis is available. However, such a flow system exists, water would have to migrate through the thick and more continuous lavas in addition to the coarse clastic units. The sedimentary units of the Springbrook and Marama Formations are discontinuous and of local extent, and hence would not make good regional aquifers. A regional flow system would also have to flow through the complex system of faults that exists in the basin. The Trout Lake Graben would present an obstacle for regional flow, as it runs perpendicular to

the suggested direction of flow, and would most likely direct the water along the various faults.

Water from Mr. Nigilser's two wells appear to be different from one another. The 178 m well may have intersected a fault or fracture zone, which acts as a groundwater conduit, giving it greater flow and different chemistry, ^2H and ^{18}O , and tritium values than the nearby 173 m well.

Chapter 5: CONCLUSIONS AND RECOMMENDATIONS

The hydrological study undertaken in the White Lake Basin has led to several conclusions. Three distinct water systems have been delineated on the basis of chemistry and isotope data. Surface waters, groundwaters in overburden and shallow bedrock aquifers, and deep groundwaters have been identified.

Most waters in the basin appear to have started out as Ca - Mg - HCO₃ type waters and evolved to their present chemical composition through one or more paths, such as normal groundwater evolution, evolution by evaporation, and cation exchange.

Local and intermediate flow systems have been suggested to exist, but chemical and geological evidence seems to rule out the presence of a large-scale regional flow system. The chemistries of the deep waters seem to be dominated by water-rock interactions in the local vicinities of the diamond drill holes.

This is in contrast to Lewis (1984) who suggests that a regional flow system exists, recharging around White Lake and flowing up dip towards hole 78-3 (northwest). At the present time, evidence from this study suggests that this is not the case.

There is no evidence that any of these waters have been subjected to significantly higher temperatures than measured at the discharge points.

More extensive sampling of the area is needed, especially in diamond drill holes. Existing holes could be used and new holes could be drilled and sampled in areas which would give a better regional picture of groundwater flow in the basin. These holes would provide data which would confirm or contradict the conclusions made in this study.

REFERENCES

- Brown, R.M., 1961. Hydrology of tritium in the Ottawa Valley. *Geochim. Cosmochim. Acta*, 22, pp. 199-216.
- Chebotarev, I.I., 1955. Metamorphism of natural waters in the crust of weathering. *Geochim. Cosmochim. Acta*, 8, pp. 22-48.
- Church, B.N., 1973. Geology of the White Lake Basin. British Columbia Ministry of Energy, Mines and Petroleum Resources, Bulletin 61, 120 p.
- _____, 1979. Geology of the Penticton Tertiary Outlier. British Columbia Ministry of Energy, Mines and Petroleum Resources. Preliminary Map 35.
- _____, Ewing, T.E. and Hora, Z.D., 1983. Volcanology, structure, coal and mineral resources of early Tertiary outliers in South-Central British Columbia. Geological Association of Canada, Joint Annual Meeting, Field Trip No. 1, 91 p.
- Clark, I.D., Fritz, P. and Michel, F.A., 1980. Isotope hydrogeology and geothermometry of the Mount Meager Geothermal Area, Final Report. British Columbia Ministry of Energy, Mines and Resources, 131 p.
- Dansgaard, W., 1964. Stable isotopes in precipitation. *Tellus*, 16, no. 4, pp. 436-468.
- Ellis, A.J. and Mahon, W.A.J., 1967. Natural hydrothermal systems and experimental hot water/rock interaction, Part II. *Geochim. Cosmochim. Acta*, 37, pp. 1255-1275.
- Fontes, J.C., 1979. Environmental isotopes in groundwater hydrology, Chapter 3. In: *Handbook of Environmental Isotope Geochemistry, Volume 1, The Terrestrial Environment*, A., Fritz, P. and Fontes, J.C. (editors), Elsevier, 1980, Amsterdam.
- Fournier, R.O. and Truesdell, A.H., 1973. An empirical Na-K-Ca geothermometer for natural waters. *Geochim. Cosmochim. Acta*, 37, pp. 1255-1275.

- Freeze, R.A. and Cherry, J.A., 1979. Groundwater. Prentice-Hall, Inc., Englewood Cliffs, N.J.
- Jessop, A.M. and Judge, A.S., 1971. Five measurements of heat flow in southern Canada. Canadian Journal of Earth Sciences, 8, no. 6, pp. 711-716.
- Kubasiewicz, M., 1983. Tritium distribution in groundwaters at a waste disposal site near Ottawa, Canada. B.Sc. Thesis, Carleton University, 45 p.
- Lewis, T.J., 1984. Geothermal energy from Penticton Tertiary Outlier, British Columbia: An initial assessment. Canadian Journal of Earth Sciences, 21, no. 2, pp. 181-188.
- Michel, F.A., 1977. Hydrogeologic studies of springs in the Central Mackenzie Valley, Northwest Territories, Canada. M.Sc. Thesis, University of Waterloo, 185 p.
- Michel, F.A. and Fritz, P., 1982. An initial study of the Hydrology of the White Lake Basin, British Columbia. University of Waterloo, 30 p.
- White, D.E., 1965. Saline waters of sedimentary rocks, In: Fluids in Subsurface Environments - A Symposium. A.A.P.G. Geological Memoir 4, pp. 342-366.