A PRELIMINARY STUDY OF THE REGIONAL GROUNDWATER FLOW IN THE MEAGER MOUNTAIN GEOTHERMAL AREA, BRITISH COLUMBIA

by

GORDON REGINALD JAMIESON

B.Sc., The University of Waterloo, 1979

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE

in

THE FACULTY OF GRADUATE STUDIES DEPARTMENT OF GEOLOGICAL SCIENCES

We accept this thesis as conforming to the required standard

THE UNIVERSITY OF BRITISH COLUMBIA

May 1981

© Gordon Reginald Jamieson, 1981
In presenting this thesis in partial fulfilment of the requirements for an advanced degree at the University of British Columbia, I agree that the Library shall make it freely available for reference and study. I further agree that permission for extensive copying of this thesis for scholarly purposes may be granted by the head of my department or by his or her representatives. It is understood that copying or publication of this thesis for financial gain shall not be allowed without my written permission.

Department of Geological Science

The University of British Columbia
2075 Wesbrook Place
Vancouver, Canada
V6T 1W5

Date June 10, 1981
ABSTRACT

The Meager Mountain geothermal area is situated 160 km north-northwest of Vancouver, British Columbia. Various types of existing and field-generated geological and hydrogeological data were employed to fully evaluate the groundwater flow regime of the study site. Mathematical modelling was carried out to determine the feasible range of regional groundwater flow characteristic in the area.

Meager Mountain is a volcano comprised mainly of andesite and dacite flows, breccias and ash. It became active in the Pliocene, fracturing the Tertiary and older granitic basement rocks through which it erupted. Subsequent alpine glaciation has deposited unconsolidated deposits of variable thickness in the valley bottom.

It can be shown that the most likely position for the water table is at an intermediate elevation in the mountain system. Apart from a very few springs at higher elevations, the discharge area is believed to be confined to the portion of the valley covered by unconsolidated deposits. Meager Creek Hotsprings and Pebble Creek Hotsprings are both located in this suggested discharge area near stream level.

Water balance calculations for the Lillooet River basin and baseflow determinations in the Meager Creek basin indicate that 14.5 to 17% of the total precipitation enters the groundwater system.
Mathematical modelling indicate that the amount of groundwater discharge is dependent on the hydraulic conductivity distribution and water table configuration but independent of the depth of the flow region. The percentage of total precipitation entering the groundwater zone is calculated to be 14-18%, correlating well with the water balance and baseflow calculations. The simulations were used to estimate the hydraulic conductivity of the various materials in the Meager Mountain system.

The representative hydraulic conductivities were found to be $10^{-2}$ to $10^{-8}$ m/s for the unconsolidated deposits, $10^{-7}$ to $10^{-4}$ m/s for the basement rock and $10^{-13}$ to $10^{-8}$ m/s for the volcanics. The vertical hydraulic conductivity may be as much as 5 times greater than the horizontal in the basement rocks. The horizontal hydraulic conductivity may be as much as 5 times greater than the vertical in the volcanics rocks. Similar vertical hydraulic conductivity values probably exist in the volcanics and in the basement.

Recommendations for future work at the Meager Mountain geothermal area include the initiation of a detailed water balance in the south reservoir area, a fracture survey of the volcanic rocks, continued mathematical modelling, and hydraulic conductivity measurements in deep drill holes.
## TABLE OF CONTENTS

ABSTRACT ......................................................... ii
LIST OF TABLES .................................................. vi
LIST OF ILLUSTRATION .......................................... vii
ACKNOWLEDGEMENTS ............................................. x

1. INTRODUCTION .................................................. 1
   Objectives .................................................... 1
   Location And Access ......................................... 4
   Previous Work .............................................. 9
      B.C. Hydro And Power Authority ......................... 12
      Energy Mines & Resources Canada ....................... 13

2. PHYSICAL SETTING ............................................. 16
   Regional Geology ............................................. 16
   Local Geology ................................................. 18
   Physiography ............................................... 23
   Thermal Springs ............................................. 25
   Cold Springs ................................................ 28

3. FUNDAMENTALS OF GROUNDWATER FLOW ...................... 31
   Darcy's Law .................................................. 31
   The Porosity Relationship With Hydraulic Conductivity ... 35
   Homogeneity And Heterogeneity Of Hydraulic Conductivity .... 36
   Isotropy And Anisotropy Of Hydraulic Conductivity ........... 38
   Water Table .................................................... 40
   Flow Nets ..................................................... 40
   Recharge Areas, Discharge Areas, And Groundwater Divides ... 42
   Steady State Flow Vs Transient Flow ....................... 43
   The Effect Of The Hydrogeologic Environment On The Groundwater Regime ........................................... 46

4. PRELIMINARY WATER BALANCE ................................ 50
   Precipitation And Temperature ............................... 51
   Evapotranspiration ......................................... 54
   Runoff ......................................................... 55
   Water Balance ............................................... 57

5. HYDRAULIC CONDUCTIVITY OF FRACTURED ROCK ........... 63
   Fracture Mapping And Data Processing Methods ............. 64
   Meager Mountain Fracture Survey ............................ 67
   Results ....................................................... 67
   Discussion .................................................... 70
   Published Fracture Permeabilities Of Various Rock Types .... 75

6. HYDROGEOLOGY ................................................. 80
   Water Table Configuration ................................... 82
   Hydraulic Conductivities Of The Geologic Materials ...... 84
# LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.1 Precipitation and temperature data at Pemberton Meadows and Bralorne</td>
<td>53</td>
</tr>
<tr>
<td>5.1 Orientation variations of joint sets</td>
<td>70</td>
</tr>
<tr>
<td>5.2 Average orientation of joint sets</td>
<td>71</td>
</tr>
<tr>
<td>5.3 Joint spacing and aperture variations</td>
<td>72</td>
</tr>
<tr>
<td>5.4 Average spacing and aperture</td>
<td>72</td>
</tr>
<tr>
<td>5.5 Summary of measured hydraulic conductivity values for various rock types</td>
<td>78</td>
</tr>
<tr>
<td>7.1 Summary of sensitivity analysis simulations</td>
<td>106</td>
</tr>
<tr>
<td>7.2 Hydraulic conductivity distributions used in FREESURF1 simulations</td>
<td>112</td>
</tr>
<tr>
<td>7.3 Recharge and hydraulic conductivity distributions with a set discharge in FREESURF1 simulations</td>
<td>124</td>
</tr>
</tbody>
</table>
# List of Illustration

<table>
<thead>
<tr>
<th>Illustration</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Location of Meager Mountain geothermal area</td>
<td>5</td>
</tr>
<tr>
<td>1.2 River, glacier, and mountain peak designation</td>
<td>6</td>
</tr>
<tr>
<td>1.3 Photograph of Meager and Plinth peaks</td>
<td>7</td>
</tr>
<tr>
<td>1.4 Hotsprings and reservoir locations</td>
<td>9</td>
</tr>
<tr>
<td>1.5 Drill hole locations</td>
<td>11</td>
</tr>
<tr>
<td>2.1 Regional geology and location of thermal springs in southwestern British Columbia</td>
<td>17</td>
</tr>
<tr>
<td>2.2 Local geology of the Meager Mountain area</td>
<td>20</td>
</tr>
<tr>
<td>2.3 Photograph of the Meager Creek valley in the South Reservoir area</td>
<td>24</td>
</tr>
<tr>
<td>2.4 Photograph of the rapidly eroding volcanics</td>
<td>24</td>
</tr>
<tr>
<td>2.5 Location of major vents at Meager Creek Hotsprings</td>
<td>27</td>
</tr>
<tr>
<td>2.6 Location of major vents at pebble creek Hotsprings</td>
<td>27</td>
</tr>
<tr>
<td>2.7 Cold spring locations</td>
<td>29</td>
</tr>
<tr>
<td>3.1 Experimental apparatus for the illustration of Darcy's law</td>
<td>32</td>
</tr>
<tr>
<td>3.2 Hydraulic head, pressure head and elevation head for a laboratory manometer</td>
<td>32</td>
</tr>
<tr>
<td>3.3 Layered heterogeneity and trending heterogeneity</td>
<td>37</td>
</tr>
<tr>
<td>3.4 Four possible combinations of heterogeneity and anisotropy</td>
<td>37</td>
</tr>
</tbody>
</table>
3.5 Relationship between layered heterogeneity and anisotropy 41
3.6 groundwater flow at various boundaries 41
3.7 Two-dimensional groundwater flow net 44
3.8 Effect of topograph on groundwater flow patterns 44
3.9 Effect of geology on groundwater flow patterns 47
4.1 Location of data gathering stations in the Lillooet River area 52
4.2 Hydrometeorological regime of the Lillooet River and valley bottom 56
4.3 Mean of mean daily discharge for Lillooet River 61
5.1 Location of fracture survey sites 68
6.1 General groundwater flow in Meager Mountain 81
6.2 Water table elevation and seepage face development 83
6.3 Equivalent hydraulic conductivity in layered volcanics 87
6.4 Cross-section of Meager Creek at stage location 87
6.5 Summary of hydrogeology on the south side of Meager Mountain 91
7.1 Region of flow for mathematical modelling 96
7.2 Line of section for region of flow 96
7.3 Geological configurations simulated 100
7.4 groundwater movement in the rock and in the unconsolidated material 102
7.5 Geometries used in FOPS simulations 105
7.6 Finite element mesh used in FREESURF1
<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.7</td>
<td>Equipotential pattern for FREESURF 1A and FREESURF 1B cases</td>
<td>110</td>
</tr>
<tr>
<td>7.8</td>
<td>Equipotential pattern for FREESURF 1C cases</td>
<td>113</td>
</tr>
<tr>
<td>7.9</td>
<td>Equipotential pattern for FREESURF 2 cases</td>
<td>114</td>
</tr>
<tr>
<td>7.10</td>
<td>Equipotential pattern for FREESURF 3 cases</td>
<td>115</td>
</tr>
<tr>
<td>7.11</td>
<td>Equipotential pattern for FREESURF 4 cases</td>
<td>116</td>
</tr>
<tr>
<td>7.12</td>
<td>Pressure vs depth graph of minimum water table elevation examples</td>
<td>117</td>
</tr>
<tr>
<td>7.13</td>
<td>Pressure vs depth graph of maximum water table elevation examples</td>
<td>118</td>
</tr>
</tbody>
</table>
ACKNOWLEDGEMENTS

I would like to express my deep appreciation to Dr. Allan Freeze for his support and guidance throughout the course of this study and for his critical review of the manuscript. I also acknowledge the support of Dr. Bill Mathews and Dr. Tom Brown for their review of this manuscript and for their participation on the thesis examination committee.

I wish to thank Dr. Jack Souther of the Geological Survey of Canada for his support and invaluable information he graciously supplied during the early stages of this study. I gratefully acknowledge the kind and helpful cooperation and assistance in the field of John Reader, Brian Fairbank, and Stu Croft of Nevin Sadlier-Brown and Goodbrande Ltd. I would like to thank Joe Stauder of the British Columbia Hydro and Power Authority for supplying my room and board in the field.

Furthermore, I would like to acknowledge the helpful advice and information supplied by Dr. Olav Slaymaker, Dr. Peter Read and Dr. Gary Clarke.

I would like to extend my thanks to Roberta Crosby for her kind assistance and counsel on the trials and tribulations of typing a thesis on the computer. The excellent drafting in this report was drawn by Gord Hodge. His work is gratefully acknowledged.
I wish to thank a good friend and colleague Grant Garven, for his helpful suggestions and assistance throughout this study and my academic program at The University of British Columbia.

A special word of thanks goes to Linda Mah, for typing the manuscript. Her constant assistance in the final preparation of this thesis has made its completion by this date a reality.
Chapter 1.
INTRODUCTION

This thesis is an evaluation of the hydrogeology of the Meager Mountain geothermal area. The work done is a small part of a joint study of Meager Mountain being carried out by the B.C. Hydro and Power Authority and Energy, Mines and Resources, Canada in order to assess the geothermal resource potential of the area.

Objectives

The main purpose of this study is to develop a preliminary mathematical model of the regional groundwater flow in the Meager Mountain geothermal area. The mathematical model will provide only an approximation of the real system, because of the simplifying assumptions that must be made in the model and the limited knowledge of the subsurface geological and hydrogeological conditions at Meager Mountain. However, it is felt that the mathematical model can prove valuable, despite the limited amount of field information. The intent here is to develop through modelling a range of groundwater flow characteristics that is both reasonable and physically possible, thereby reducing the infinite range of possibilities to a suite of feasible flow fields. As more groundwater information becomes available in the future the suite can be further refined.

It must be stressed that this is a regional study of
groundwater flow in the entire mountain and not a geothermal resource study at a local scale. Before one can focus on a specific area, the system as a whole must be understood.

The incorporation of heat flow in the groundwater model is beyond the scope of this report. Heat flow would slightly alter the subsurface flow paths but no substantial change in the general flow characteristics would be expected. It is suggested that future mathematical modelling should couple the heat flow with the groundwater flow.

In Chapter 1, the previous work undertaken in the Meager Mountain area by the British Columbia Hydro and Power Authority and the Geological Survey of Canada will be discussed. Numerous geological, geochemical and geophysical surveys have been completed since 1973.

In Chapter 2, the physiography and the complex local and regional geology are examined. The physiographic nature and location of the Meager Mountain hot and cold springs are also considered.

In Chapter 3, the basic physics of groundwater flow is discussed. The material is covered in greater detail than is normal in hydrogeological reports in an attempt to generate an understanding of groundwater flow across the interdisciplinary boundaries of the Meager Mountain geothermal project.

In Chapter 4, the climatic conditions and the amount of runoff out of the Lillooet River basin are discussed and a
calculation of a preliminary water balance for the basin is made. The objective is to estimate the percentage of the total precipitation that enters the groundwater system. The characteristics of the subsurface flow regime are dependent on the amount of water entering the system.

In Chapter 5, special attention is paid to the hydraulic conductivity of fractured rock. Fracture mapping and data processing methods are considered. The results and significance of the Meager Mountain fracture survey are discussed. To check the validity of the survey, the results are compared with published data on the fracture permeability of various rock types.

In Chapter 6, aspects of the hydrogeology are examined including the water table configuration and hydraulic conductivity of the different geologic materials. The amount of groundwater recharge in the Meager Creek basin is calculated and compared with the amount determined by the water balance, for the entire Lillooet River basin.

In Chapter 7, a discussion of the groundwater modelling is presented to illustrate that a shallow flow system on the order of 2 km deep or less, rather than a deeper-seated system, is more likely to produce the total discharge of water observed in the mountain discharge areas.
Location and Access

The Meager Mountain volcanic complex is located 160 km north-northwest of Vancouver, British Columbia and approximately 60 km northwest of Pemberton (Fig. 1.1). Most of the mountain complex and the area of concern lie south of the Lillooet River and north of Meager Creek (Fig. 1.2).

Access to the mountain complex is by paved highway from Vancouver to 20 km northwest of Pemberton where gravelled Forestry Development and private logging roads continue to the mountain area. Access within the area is by logging road or recently built by B.C. Hydro roads (Fig. 1.2).

Meager Mountain has a number of volcanic peaks including Plinth Peak, Mount Meager, Mount Job, Capricorn Mountain and Pylon Peak. Figure 1.3 is a view of the Meager and Plinth peaks from the Lillooet River valley. The mountain area is capped by an extensive system of glaciers that seasonally feeds many of the high gradient streams on the mountain side. The streams feeding Meager Creek include Devastation Creek, Boundary Creek, No Good Creek, Angel, Camp, Canyon and Capricorn Creek. Mosaic Creek, Affliction, Job and Fall Creek all feed into the Lillooet River off the north and east flanks of the mountain.

The major hotsprings in the areas are the Meager Creek and Pebble Creek Hotsprings (Fig. 1.4). The Meager Creek Hotsprings are approximately 6 km from the confluence of
Figure 1.1 Location of Meager Mountain (after Lewis and Souther, 1978)
Figure 1.2 River, glacier and mountain peak names in the Meager Mountain area.
Figure 1.3 Meager and Plinth peaks.
Meager Creek with the Lillooet River. The Pebble Creek Hotsprings are located on the northwest side of the Lillooet River, 7.5 km upstream from its intersection with Meager Creek.

The two most promising areas for geothermal development, as determined by resistivity surveys, are outlined in Figure 1.4. The region along the Lillooet River including the Pebble Creek Hotsprings has been designated the North Reservoir. The area north of Meager Creek including No Good and Angel Creek is called the South Reservoir.

In geology, the term reservoir indicates there has been an accumulation of fluid in a permeable geologic unit under adequate trap conditions. Trap conditions are not essential to explain the presence of hot water in the basement rock. A more plausible explanation is that regional groundwater flow through the more permeable zones or aquifers in the rock is supplying a constant flow of hot water to the area. This explanation will become clear later in the report. To be consistent with previous reports the terms North and South Reservoir will be used, to refer to the two areas as they are outlined in Fig. 1.2.

Most of the field work discussed in the next section has been concentrated on the South Reservoir due to its easier accessibility and its more promising, initial geophysical and drilling results.
Figure 1.4 Hotsprings and reservoir locations in the Meager Mountain area.
Previous Work

In 1973 the Department of Energy, Mines and Resources Canada through the Geological Survey of Canada and the Earth Physics Branch began to assess the geothermal resource potential of western Canada. The location and age of Quaternary volcanics and high level plutons were catalogued and this study was followed by a geochemical survey of thermal springs to identify the waters most likely to have been at much higher temperatures at depth. As a result of this work the Meager Mountain area was chosen as the most favourable region for more detailed investigation (Lewis and Souther, 1978).

Since 1973 various surveys in the Meager Mountain area have been conducted by the Geological Survey of Canada and the Earth Physics Branch of Energy Mines and Resources Canada and by Nevin, Sadlier-Brown, Goodbrande Ltd., and their subcontractors for the British Columbia Hydro and Power Authority. Much of the early work involved geophysical studies such as resistivity, self-potential, seismic and magnetotelluric surveys. Extensive geologic mapping of the region and geochemistry of the thermal and cold waters has been completed. In recent years a greater amount of drilling has been undertaken in the promising areas delineated by the initial geophysical work. The location of these drill holes is illustrated in Figure 1.5.

The following summary of work completed is divided into two lists, one for each of the two main participants in the
Figure 1.5 Drill hole locations in the Meager Mountain area.
project. A portion of this summary is taken from Fairbank et al. (1979).

B.C. Hydro and Power Authority

1974 Geological, geochemical and geophysical surveys were initiated.

1975 Dipole-dipole resistivity surveys, diamond drilling and water geochemistry studies were performed on the south side of the mountain area in the Meager Creek valley. This work defined the South Reservoir as a tabular shaped body open to the north under Meager Mountain.

1976 A self-potential(SP) geophysical survey on the north side of the mountain was inconclusive.

1977 A resistivity low was delineated near the confluence of the Lillooet River and Pebble Creek using a pole-pole resistivity survey.

1978 Pole-pole resistivity work on the north and south side of the mountain further delineated the reservoir regimes. A minor dipole-dipole survey was undertaken in the North Reservoir area. Two exploratory holes were drilled, one in the North Reservoir and one in the South Reservoir. Twelve percussion drill holes were sunk along Meager Creek and near the Lillooet River approximately 1 km north of its confluence with Meager Creek. Geologic mapping was accomplished, mostly in the South Reservoir area. Radon gas and
mercury surveys were undertaken in an attempt to de-
lineate geothermal water pathways to the surface.

1979 The dipole-dipole resistivity survey was continued in
the North Reservoir area. Three exploration holes
were drilled in the South Reservoir. Geologic map-
ing of the South Reservoir area was continued. A
reconnaissance study of basement-rock alteration was
initiated to determine if there is a hydrothermal
alteration pattern associated with the inferred
north-south structural zone in either the North or
South Reservoir (Fairbank et al., 1980).

1980 Five exploration holes were drilled in the South
Reservoir. Geologic mapping was continued in the
North and South Reservoir areas. A fracture survey
of the basement granodiorites was undertaken in the
South Reservoir.

Energy Mines & Resources Canada

1973 Two 50 m diamond drill holes were bored at the Meager
Creek Hotsprings.

1974 Microseismicity studies were completed in the Meager
Mountain area.

1976 Seismic profiling was undertaken in the upper Lillooet
Valley, magnetotelluric surveys were executed between
Meager Creek and Pemberton Meadows in the Lillooet
Valley. Diamond drilling and studies of temperature
gradients were undertaken in the Lillooet and
Squamish valleys by Lewis (1977) of the Earth Physics Branch. A water geochemistry study of Pebble and Meager Creek Hotsprings and surface waters was done by Hammerstrom and Brown (1977) at the Department of Geological Sciences, University of British Columbia and found no evidence that the water had been heated above 80°C.

1977 Detailed geological and stratigraphic mapping of the Meager Mountain volcanic complex was initiated by Read (1977) for the Geological Survey of Canada. Michel and Fritz (1977) of the University of Waterloo, Department of Earth Science performed an isotope study of stream water, spring water and snow samples in the Meager Mountain area to interpret the origin, history, flow and chemistry of the natural discharging groundwater.

1978 Read continued to map the volcanic complex. Lewis and Souther (1978) released a summary and interpretation of the information obtained at Meager Mountain to date with respect of its geothermal resource potential.

1979 Read (1979) continued mapping and released a geological map of the Meager Mountain area. Clark (1980) of the University of Waterloo, Dept. of Earth Sciences, undertook an isotope hydrogeology and geothermometry study of the thermal waters at Meager Mountain and found no evidence that the waters were heated above
140°C.

With the familiarizing introductory material at hand, the physical setting of the Meager Mountain area can now be described in detail.
Chapter 2.

PHYSICAL SETTING

In an attempt to interpret the hydrogeology of an area one must first have an understanding of the material through which the water is flowing and of the locations where the water is discharging from the material. It is therefore essential that the geology of an area and the spring locations are known, if reasonable hydrogeological interpretations are to be made. This chapter discusses the geology of the Meager Mountain area and the physical setting of the hot and cold springs.

Regional Geology

Meager Mountain is situated in the Coast Mountains near the axis of the Coast Plutonic Complex, a northwesterly trending belt of Tertiary and older granitic and metamorphic rocks. The northwesterly trending Pemberton Belt of late Tertiary and Quaternary plutons, and the Garibaldi Belt of north-south trending Quaternary volcanoes, intersect within the Coast Plutonic Complex at the Salal Pluton near Meager Creek (Fig. 2.1) (Lewis and Souther, 1978). Potassium-argon dates suggest that the plutons of the Pemberton Belt range from 7.9 to 18 Ma old. The volcanoes of the Garibaldi Belt are much younger, giving potassium-argon ages of 4 Ma to less than 100,000 years. The Pemberton Belt is believed to consist of subvolcanic roots of a Miocene volcanic front related
Figure 2.1 Regional geology and the location of thermal springs in southwestern British Columbia (after Lewis and Souther, 1978)
to the subduction of the Juan de Fuca Plate (Lewis and Souther, 1978). The six andesite-dacite volcanoes of the Garibaldi Belt are also considered to be related to this subduction, and are believed to be an extension of the High Cascades in the western United States, which include Mount Baker, Mount St. Helens, and other volcanic centres (Clark, 1980).

In the Meager Mountain area the Coastal Plutonic complex consists of northwest trending discontinuous strips of metavolcanics and metasediments surrounded by quartz diorites and granodiorites and all overlain by isolated patches of younger volcanic rocks. The Salal Creek Pluton, a quartz monzonite, cuts the older plutonic and metamorphic rocks near the head of the Lillooet River. Some smaller satellite bodies of the pluton underlie part of the north portion of Meager Mountain. The Salal Creek Pluton is part of the Pemberton Belt of high level plutons.

Local Geology

The Meager Mountain Volcanic Complex was originally mapped by Anderson (1975) for a B.Sc. thesis at the University of British Columbia. More recently Read (1977, 1979) mapped the area in greater detail under contract to the Geological Survey of Canada.

The older portion of the complex comprises mainly widespread andesite and is best exposed in the south. The younger
north half is composed of dacite flows and lava domes overlying the older andesite flows.

Read (1979) initially broke the complex into nine volcanic assemblages. With further mapping and age dating Read (1979) subdivided the complex further into 18 volcanic assemblages, four in the Pliocene and 14 in Quaternary time. Lewis and Souther (1978) grouped the assemblages of Read (1977) into four main units or phases which will be discussed below.

The bottom of the volcanic pile consists of basal breccia which has incorporated blocks of basement in a tuffaceous matrix, exceeding 300 m in thickness along the southern side of the complex (Read, 1977). This indicates that the initial eruption was explosive leading to extensive fracturing of the local basement rock. Directly overlying the breccia is a sequence consisting of dacite flows and up to 500 m of acid tuff. The breccia, flows and tuff are all part of unit 1 (Fig. 2.2).

The porphyritic andesites and minor hypabyssal intrusions of unit 2 make up the main mass of the complex. Potassium-argon dates reported by Lewis and Souther (1978) range from 4.2 ± 0.3 Ma to 2.1 ± 0.2 Ma suggesting a long period of intermittent andesitic volcanism.

The youngest rocks of the complex, unit 3, are dacite flows, breccias, tuffs and hypabyssal intrusives. The unit
Figure 2.2. Geology of the Meager Mountain area (after Lewis and Souther, 1979).
is up to 600 m thick. The Meager, Capricorn, Job and Plinth summits along with a large portion of the central and north-east part of the complex are comprised of unit 3.

The most recent volcanic activity at Meager Mountain occurred 2440 ± 140 years ago with the discharge of the Bridge River ash (Nasmith et al., 1967) which covered an area as far east as Banff Park, Alberta. Deposits of the ash near the mountain are up to 30 m thick and constitute the youngest portion of the Bridge River unit.

The Meager Creek and Lillooet valleys, surrounding Meager Mountain, have been filled with varying thicknesses of unconsolidated glacial deposits. In Lillooet valley drill hole L1-78D (Fig. 1.4) encountered 47 m of outwash sands and gravels with interbedded bouldery till layers. The drill hole is situated near the contact between the valley fill and the rock slopes. The unconsolidated deposits are thought to be much deeper closer to the river axis.

Drill hole M5-78D in the Meager Creek valley was sunk 250 m and failed to intersected bedrock. Outwash sands and gravels, interbedded zones of large boulders and rare thin glaciolacustrine clay layers were encountered. Farther down stream at drill hole EMR73-1 and EMR73-2 the overburden thins to 18 m or less, partly because of a bedrock high and valley narrowing in the area.

The granitic basement in the Meager Mountain area is
highly jointed and fractured because of the explosive nature of the initial Meager Mountain eruptions. A fracture survey revealed that there are 2 dominant joint sets and two subordinate sets in the South Reservoir area. More details of the fracture survey will follow in Chapter 5.

The volcanic rocks are also highly jointed. An extensive fracture survey was not carried out in the volcanics but steeply dipping sets appear to dominate.

The groundwater movement in the basement and volcanic rocks will be mainly along the fractures in the rock. The amount of water that moves through the relatively impermeable rock matrix should be essentially negligible. In the volcanic rocks the fracture patterns are therefore much more important than the lithology with respect to groundwater movement. In the hydrogeologic modelling which is described in Chapter 7, the volcanics are lumped into one hydrogeologic unit.

Before describing the physiography of the hot and cold springs at Meager Mountain, a brief description of the basic physiography of the area is in order.
Physiography

The impressive geology of the Meager Mountain area is surpassed only by its spectacular physiography.

The relief at Meager Mountain is more than 2300 m with the elevation ranging from 425 m to over 2700 m above sea level. The topography is very rugged with elevation changes of more than 2000 m in less than 5 km horizontal distance. The Lillooet River valley and the Meager Creek valley range in elevation from 425 m at their confluence to 900 m in their upper reaches. Figure 2.3 depicts a portion of the Meager Creek valley in the South Reservoir area. The mountain peaks including Plinth, Pylon, Meager, Capricorn and Job range between 2450 to 2700 m in elevation. The ruggedness of Meager and Plinth peaks is illustrated in Figure 1.3.

The steep slopes and the low resistance of the volcanics to weathering, along with the high precipitation in the area, leads to very rapid erosion of the volcanic pile. Figure 2.4 illustrates one of the many deeply incised valleys and canyons disintegrating at a rate too rapid for vegetation to take hold.

The mountain is capped by an extensive glacier system (Fig. 1.2) that terminates at the headwaters of a number of streams. Many of the glaciers have acquired the same name as the streams to which they supply meltwaters, such as Mosaic, Affliction, Devastation, Job and Capricorn glaciers.
Figure 2.3 Meager valley in the south reservoir area.

Figure 2.4 Rapid erosion of the volcanic rock.
The gradient of the streams on Meager Mountain is extremely high at approximately 0.3 to 0.4. The high gradient and deep, steep-sided valleys are characteristic of streams in their young stage of development. The stream pattern (Figure 1.4) from Devastation Creek counter clockwise to Manatee Creek, is a radial pattern, typical of a volcanic mountain.

Fir, hemlock and cedar trees dominate the subalpine forest below 1500 to 1800 m. Alpine meadows exist above the tree line, in areas were erosion is slow enough to allow for their development.

**Thermal Springs**

The hot springs of southwestern British Columbia can be divided into two groups. The first group is associated with the Pemberton Belt (Fig. 2.1) and includes well known hot springs such as Harrison and Sloquet. It appears that the fault system of the Pemberton Belt has supplied the fracture permeability needed to allow for the deep circulation of meteoric waters and their consequent heating.

The second group of hot springs are associated with Quaternary volcanism in the Meager Mountain and Mount Cayley areas (Fig. 2.1). The two hotspring areas of interest to this research are the Pebble Creek hotsprings and Meager Creek hotsprings located at Meager Mountain (Fig. 1.2, 2.2).
Meager Creek hotsprings issue from coarse fluvial sand and gravel deposits approximately 6 km from the confluence of Meager Creek with the Lillooet River. More than 30 springs and seeps issue from a 1200 square meter area (Fig. 2.5) with a total discharge of approximately 40 l/s, at temperatures of 45-55°C (Lewis and Souther, 1978).

It is believed that the location of the hotsprings is controlled by an underlying bedrock topographic high. In Figure 1.4 the outline of the South Reservoir shows a tongue extending down the Meager Creek valley to the Meager Creek Hotsprings. Thermal waters probably enter the valley fill, travel down gradient and exits at the present hot springs site. The rise in the bedrock causes a thinning of the valley fill, which in turn, gives rise to hot water surfacing at the present position of the Meager Creek Hotsprings.

One kilometer upstream from Meager Creek Hotsprings is the Placid Hotsprings, discharging from a gravel bank of Meager Creek at an estimated rate of 2 l/s at 45°C. Five kilometers farther upstream is the No Good Warm Springs issuing 20-40°C water from a grassy sandy bank on the north side of Meager Creek at a rate of 5 l/s. The warm springs represent the most westerly discharge of thermal waters found at Meager Creek.

The Pebble Creek Hotsprings are located on the northeast side of the Lillooet River, 7.5 km upstream from its confluence with Meager Creek. The two main vents occur on a 20 m
Figure 2.5 Location of major vents at the Meager Creek hotsprings site (after Clark, 1980).

Figure 2.6 Location of major vents at the Pebble Creek hotsprings site (after Clark, 1980).
high bench, situated on the bank of the river (Fig. 2.6). The thermal waters deposit calcite tufa, stained deep ochre in the area of the vents. Additional seeps issue from the quartz monzonite bedrock, unconsolidated deposits and pyroclastic debris outcropping on the face of the bench near the river.

The Pebble Creek Hotsprings issue directly from bedrock, unlike the Meager Creek Hotsprings which issue from valley fill.

**Cold Springs**

A number of cold springs exist in the Meager Mountain complex, ranging in altitude from 580 m to 1850 m. Figure 2.7 shows the location of the major cold springs discovered to date. The springs are found in the basement rock, volcanic rock and the unconsolidated material.

The sites of springs in the unconsolidated material are generally controlled by the stratigraphy of the deposits. The springs and seeps are usually found at contacts between silt or clay and coarse alluvium exposed on the steep sides of V-shaped stream valleys. The water travelling in the permeable alluvium intersects the clay layer and cannot penetrate it. The water is, therefore, forced to travel along the contact and issue as a spring.

A number of cold springs at high elevations discovered
Figure 2.7 Location of the cold springs in the Meager Mountain area.
by Read (1979) and Clark (1980) are also stratigraphically controlled. The springs discharge at geologic contacts between volcanic layers, volcanic-basement contacts and volcanic-unconsolidated-deposit boundaries.

The author, during his fieldwork, did not observe all 14 cold springs marked on Figure 2.7. The springs investigated had discharge rates of 2 to 25 1/s. Calculations, that will be discussed in Chapter 6, show that the cold springs contribute a maximum of only 2 to 3 percent of the total groundwater discharge observed in the Meager Creek Basin. The amount of groundwater released by the cold springs is therefore, thought to be insignificant.

It should be noted that most of the cold springs and all of the hot springs exist in local, low-elevation areas near creeks and streams. In the following chapter, regional groundwater flow systems will be discussed. It will be shown that water enters the system at higher elevations and discharges out of the system in the low-lying areas. The springs naturally occur in the low-lying discharge areas.

Now that the physical setting has been discussed, the hydrologic and hydrogeologic aspects of the Meager Mountain area can be examined. Before this examination, background information of the basic physics of groundwater flow will be considered in Chapter 3.
Chapter 3.
FUNDAMENTALS OF GROUNDWATER FLOW

This chapter discusses the basic physics of groundwater flow in an attempt to generate an understanding of hydrogeologic environments across the interdisciplinary boundaries of the Meager Mountain geothermal project. Readers with a hydrogeological background may want to move directly to Chapter 4. The majority of the material in this chapter was taken from Freeze and Cherry (1979). The interested reader may wish to refer to this text if more detail is desired.

Darcy's Law

The science of groundwater hydrology began in 1856 when Henry Darcy published a report on his laboratory experiment analyzing the flow of water through sands. Darcy's experiment was set up as in Figure 3.1. A circular cylinder of cross section A is filled with sand, stoppered at each end, and equipped with inflow and outflow tubes and a pair of manometers. Water is allowed to flow through the sand until the inflow rate Q is equal to the outflow rate. The elevations of the manometer intakes are \( z_1 \) and \( z_2 \) and the elevations of the fluid levels are \( h_1 \) and \( h_2 \) with respect to an arbitrary datum set at \( z=0 \). The distance between the manometer intakes is \( \Delta l \). The difference in fluid levels \( h_1 - h_2 \) is denoted as \( \Delta h \).

Let us define \( v \), the specific discharge through the cy-
Figure 3.1 Experimental apparatus for the illustration of Darcy's law (after Freeze and Cherry, 1979).

Figure 3.2 Hydraulic head $h$, pressure head $\psi$, and elevation head $z$ for a laboratory manometer (after Freeze and Cherry, 1979).
\[ v = \frac{Q}{A} \quad (3.1) \]

where \( v \) has the dimensions of \([L/T]\) if the dimensions of \( Q \) are \([L^3/T]\) and those of \( A \) are \([L^2]\). The SI units for \( v \) would be m/s. Although \( v \) has the dimensions of velocity it is better thought of as a flux rate \([L^3/T/L^2]\).

Darcy's experiment showed that \( v \) is directly proportional to \( \Delta h \) and inversely proportional to \( \Delta l \). Darcy's law in one dimension can then be written as,

\[ v = -K\frac{\Delta h}{\Delta l} \quad (3.2) \]

or, in differential form,

\[ v = -K\frac{dh}{dl} \quad (3.3) \]

where \( K \) is a constant of proportionality.

In equation 3.3, \( K \) is known as the hydraulic conductivity, \( h \) is called the hydraulic head, and \( \frac{dh}{dl} \) is the hydraulic gradient. Below is a detailed explanation of these parameters.

Freeze and Cherry (1979), quoting Hubbert (1940), define a potential as "a physical quantity, capable of measurement at every point in a flow system, whose properties are such that flow always occurs from regions in which the quantity has higher values to those in which it has lower, regardless of the direction in space." They note that a potential should have dimensions of energy per unit mass. Through the use of elementary physics, Hubbert illustrated that the fluid
potential $\phi$ for groundwater flow at any point in a porous media is simply the hydraulic head multiplied by the acceleration due to gravity

$$\phi = gh \quad (3.4)$$

Since $g$ is near constant, the hydraulic head $h$ is just as suitable a potential as $\phi$, and hydrogeologists find it easier to work with. In the field, the hydraulic head can be measured by installing a piezometer which is sealed along its length and allows water to enter only at a single point. The hydraulic head is given by the elevation of the water level in the piezometer.

It can be shown the hydraulic head has two components, the elevation (gravity) term and fluid pressure term. If a sand-filled cylinder were set up vertically, fluid would flow through it in response to gravity alone. On the other hand, if the cylinder were horizontal, gravity would play no role and to induce flow the pressure at one end must be greater than the other (Fig. 3.2). Therefore,

$$h = z + \psi \quad (3.5)$$

where $z$ is the elevation head above an arbitrary datum and is the pressure head. Both $\psi$ and $z$ have dimensions [L] and are commonly measured in meters. The pressure head, $\psi$, is related to the fluid pressure $p$, by

$$p = \rho g \psi \quad (3.6)$$

where $\rho$ is the density of the fluid. Piezometer measurements can therefore provide both hydraulic heads and fluid pressures.
The constant of proportionality in Darcy's law, the hydraulic conductivity $K$, is a function not only of the porous media but also the fluid. Experimentation has found the relationship to be

$$K = \frac{k \rho g}{\mu}$$

(3.7)

where $k$ is the specific or intrinsic permeability, $\rho$ is the density of the fluid, $g$ is the acceleration due to gravity, and $\mu$ is the dynamic viscosity of the fluid. The permeability $k$ or capacity for transmitting a fluid is a function of the medium only and has dimensions $[L^2]$. The density and viscosity are properties of the fluid.

The Porosity Relationship with Hydraulic Conductivity

If the total unit volume $V_r$ of a rock or soil is partitioned into the volume of the solids $V_s$, and the volume of the voids $V_v$, the porosity $n$ is defined as

$$n = \frac{V_v}{V_r}$$

(3.8)

It is usually reported as a percentage or a decimal fraction.

The porosity can be an important controlling influence on the hydraulic conductivity $K$. In general, an increase in porosity increases the permeability $k$. The permeability in turn increases the hydraulic conductivity, if fluid properties are constant.

In rock, two types of porosity exist, the intergranular porosity and the fracture porosity. The intergranular porosity refers to the porosity of the rock matrix. In a grani-
tic rock this is to all intents and purposes negligible. The fracture porosity refers to the percentage of the total volume that is taken up by the fractures in the rock. The fracture porosity may be as large as 1 or 2 percent or as small as $10^{-3}$ to $10^{-4}$. The interconnective structure of the fractures will influence the hydraulic conductivity more than the intergranular porosity.

In the fractured volcanic rocks and basement rocks of the Meager Mountain area the fracture porosity is the major controlling factor on the hydraulic conductivity. In the granular unconsolidated deposits intergranular porosity is the controlling factor.

**Homogeneity and Heterogeneity of Hydraulic Conductivity**

If the hydraulic conductivity $K$ is independent of position within a geologic formation, the formation is homogeneous. If the hydraulic conductivity $K$ is dependent on position within a geologic formation, the formation is heterogeneous. If we set up an $xyz$ coordinate space, then in a homogeneous formation, $K(x,y,z)=C$, $C$ being a constant; whereas in a heterogeneous formation $K(x,y,z)=C$ (Freeze and Cherry, 1979).

A variety of environments can cause heterogeneity. A system may be heterogeneous due to layering (Fig. 3.3) such as in the volcanics of Meager Mountain. An individual flow or breccia layer may have a homogeneous hydraulic conductivi-
Figure 3.3 Layered heterogeneity and trending heterogeneity (after Freeze and Cherry, 1979).

Figure 3.4 Four possible combinations of heterogeneity and anisotropy (after Freeze and Cherry, 1979).
ty but when one looks at the whole volcanic pile, the system is heterogeneous. A discontinuous heterogeneity can be caused by a fault or by a large-scale stratigraphic features such as the overburden-bedrock contact. Within a formation one can also get trending heterogeneity (Fig. 3.3). Trends in a volcanic layer in the Meager Mountain complex could occur in flows partially exposed at surface. Fracture apertures would be wider at the surface than in unexposed sections causing a decrease in hydraulic conductivity with depth.

Isotropy and Anisotropy of Hydraulic Conductivity

A formation is isotropic if its hydraulic conductivity $K$ at any point is of the same magnitude in all directions. A formation is anisotropic if, on the other hand, the hydraulic conductivity is affected by the choice of direction of measurement at a point (Fig. 3.4) (Davis and De Wiest, 1966).

If one deals with the three-dimensional case there will be three principal directions of hydraulic conductivity $K_x$, $K_y$, and $K_z$. At any point in an isotropic formation $K_x = K_y = K_z$, whereas in an anisotropic formation $K_x = K_y = K_z$.

The basement granodiorite at Meager Mountain appears to be anisotropic. The two prominent joint sets have a near vertical dip causing the vertical hydraulic conductivity $K_z$ to be greater than the horizontal hydraulic conductivity $K_x$. 
The volcanic rocks, on the other hand, tend to be a system of interbedded high-permeability breccias and ash layers and lower-permeability flows. It can be shown that there is a relationship between such layered heterogeneity and anisotropy. Figure 3.5 is a layered formation where each layer is homogeneous and isotropic with a hydraulic conductivity \( K_1, K_2, \ldots, K_n \). It can be shown that an equivalent vertical hydraulic conductivity \( K_z \) for the system of layers can be calculated from the relation

\[
K = \frac{d}{\sum_{i=1}^{n} \frac{d_i}{k_i}}
\]

where \( n \) is the total number of layers, \( d_i \) is the total thickness of the \( i \)th layer, and \( k_i \) is the hydraulic conductivity of the \( i \)th layer.

Similarly, it can be shown that the equivalent horizontal hydraulic conductivity \( K_x \), for the layered system is

\[
K = \sum_{i=1}^{n} \frac{k_i d_i}{d}
\]

With some mathematical manipulation of Equation 3.9 and 3.10 it is possible to show that \( K_x > K_z \) for all possible values of \( K_1, K_2, \ldots, K_n \). In the field, it is not uncommon for layered heterogeneity to lead to regional anisotropy values on the order of 100:1 or even greater. In the layered volcanics at Meager Mountain the anisotropy is reduced due to extensive vertical and near vertical fracturing that in-
creases the vertical hydraulic conductivity $K_z$.

**Water Table**

The water table is defined as the surface on which the fluid pressure $p$ in the pores of the medium is exactly atmospheric. In gauge pressure, $p$ would be equal to zero, implying from Equation 3.6 that $\psi = 0$. Since $h = \psi + z$, the hydraulic head at any point on the water table must be equal to the elevation $z$ of the water table at that point. Alternatively, the water table can be viewed as an imaginary surface below which the pore spaces are completely saturated with water.

**Flow Nets**

If hydraulic head values are known throughout a two-dimensional system, contours can be drawn joining up points of equal potential. These contours are called equipotential lines. Flowlines can be constructed perpendicular to the equipotential lines which is the direction of the maximum potential gradient. Flowlines indicate the direction of water movement. The resulting set of flowlines and equipotential lines is known as a flow net.

When constructing a flow net for a homogeneous, isotropic system with saturated, steady-state flow, three types of boundaries can exist: 1) impermeable boundaries, 2) constant head boundaries and 3) water table boundaries. Flow in the vicinity of an impermeable boundary [Figure 3.6(a)] must be
Figure 3.5 Relationship between layered heterogeneity and anisotropy (after Freeze and Cherry, 1979).

Figure 3.6 Groundwater flow in the vicinity of (a) an impermeable boundary, (b) a constant head boundary, and (c) a water-table boundary (after Freeze and Cherry, 1979).
parallel to the boundary because no flow can move across it. This demands that flow lines are parallel to the boundary and equipotentials are at right angles (Freeze and Cherry, 1979).

In the case of a constant head boundary, flow must meet the border at right angles because the boundary is actually an equipotential line [Figure 3.6(b)].

As discussed previously, at the water table the hydraulic head \( h \) is equal to the elevation \( z \). The water table is consequently neither a flow line or an equipotential line [Figure 3.6(c)].

Flow nets can be constructed for any of the systems illustrated in Figure 3.4 and used to calculate the discharge through the system. For a thorough explanation of flow net construction and discharge calculations see Freeze and Cherry (1979) pp. 168-189.

**Recharge Areas, Discharge Areas, and Groundwater Divides**

Consider the two-dimensional cross section Figure 3.7 of a set of parallel ridges and valleys with an impermeable boundary at the base. The geologic materials are homogeneous and isotropic. The water table configuration is a subdued replica of the topography in the hills and coincident with the ground surface in the valleys. This water table shape is characteristic of most topography, although the author is unaware of any specific studies of the water table configura-
tion in mountainous terrain. The hydraulic head value on any one of the dashed equipotential lines is equal to the elevation of the water table at its point of intersection with the equipotential line (Freeze and Cherry, 1979).

From Figure 3.7, it is clear that, groundwater flows from the highlands towards the valleys. The symmetry of the system produces vertical boundaries at AB and CD under the valleys and ridges, and these boundaries are known as groundwater divides. It is obvious no flow enters or leaves the area ABCD through the lines AB or CD, thereby indicating they are "imaginary" impermeable boundaries.

The region ED in Figure 3.7 is known as the recharge area. In a recharge area the net component of saturated groundwater flow is downward, away from the water table. The region AE in Figure 3.7 is known as the discharge area. In a discharge area the net component of saturated groundwater flow is upward, towards the water table. A cross section of the Meager Mountain area, similar to ABCD will be used for mathematical modelling purposes in Chapter 7.

**Steady State Flow vs Transient Flow**

Steady-state flow takes place when at any point in a flow field the direction and magnitude of the flow velocity are constant with time. Transient flow, also known as nonsteady or unsteady flow, takes place when at any point in a flow field the direction and magnitude of the flow velocity
Figure 3.7 Groundwater flow net in a two-dimensional vertical cross-section through a homogeneous, isotropic system bounded on the bottom by an impermeable boundary (after Hubbert, 1940).

Figure 3.8 Effect of topography on regional groundwater flow patterns (after Freeze and Witherspoon, 1967).
changes with time.

Flow nets can be drawn for both steady-state and transient flow. A flow net for a steady-state system represents the system at all times. A flow net for a transient system only represents the system for a particular instant in time.

In regional groundwater flow, steady-state flow can be thought of as representing the hypothetical situation where the water table maintains the same position throughout the entire year. In reality, fluctuations in the water table throughout the year introduce transient effects into the flow system. However, if the changes in the water table position are small compared with the total thickness of the system, and if the relative configuration of the water table stays the same throughout the fluctuations, we can replace the fluctuating system with a steady-state system with the water table fixed at its mean position (Freeze and Cherry, 1979).

In a steady-state system, no water in the system is given up from storage in the medium and no water is taken up into storage in the medium. Therefore, the total recharge into the system must equal the total discharge out of the system. The regional groundwater flow in Meager Mountain is assumed to be steady-state for the purpose of hydrogeologic interpretation and mathematical modelling.
The Effect of the Hydrogeologic Environment on the Groundwater Regime

The groundwater regime is controlled by a number of parameters which collectively create the hydrogeologic environment of a given geographic region. Among the most important parameters are topographic relief, rock porosity and hydraulic conductivity, precipitation and evapotranspiration.

Minor changes in topography can vary the groundwater flow a great deal. Figure 3.8 depicts two cross sections having an upland area to the right and a valley to the far left. Figure 3.8(a) shows a flat upland and water table while Figure 3.8(b) illustrates a hilly upland area with a hummocky water table configuration. The difference between the flow systems is obvious. A few small hills change a single flow pattern into numerous subsystems within a major flow system. Clearly, even basins underlain by homogeneous, isotropic geologic materials can have complex systems of groundwater flow due to topography alone (Freeze and Cherry, 1979).

Figure 3.9 shows the effect of different geologic environments on groundwater flow as mathematically simulated by Freeze and Witherspoon (1967). Figure 3.9(a) and 3.9(b) represent an aquifer at depth with a conductivity 10 and 100 times that of the overlying formation. Water enters the system, flows nearly vertically downward to the aquifer, travels horizontally in the aquifer and finally upwards in the dis-
Figure 3.9 Effect of geology on regional groundwater flow patterns (after Freeze and Witherspoon, 1967).
charge area.

In Figure 3.9(c) the aquifer at depth acts as a through-way for flow to pass under the overlying local systems.

An aquifer that pinches out creates a discharge area in the centre of the section in Figure 3.9(d).

Figure 3.9(e) illustrates how the difference of only a few meters in the point of recharge can determine whether water enters a small local flow system or large regional flow system.

It is clear from these simple examples that the complex geology of the Meager Mountain area can have an enormous effect on the groundwater flow pattern.

Other factors affecting the groundwater flow such as precipitation, and evapotranspiration will determine the amount of water available to reach the water table in a recharge zone. High precipitation with low evapotranspiration would allow a large quantity of water to flow through the unsaturated zone to the water table, raising the water table closer to the surface. Conversely, in an area with low precipitation and high evapotranspiration one would expect the water table elevation to be lower than in the first case.

Groundwater flow is an extremely dynamic process very dependent on its hydrogeologic environment.

This review of groundwater fundamentals has been pre-
sented to enable non-hydrogeologists to gain a better understanding of the hydrogeology of the Meager Mountain complex as it is presented in Chapters 6 and 7.

Firstly, however, it is necessary to place the groundwater regime into the context of the complete hydrologic cycle. In the following chapter we will develop a water balance for the Lillooet River basin. In Chapter 7, it will be shown that this water balance and the hydrogeological model provide a consistent picture of the hydrology of the area.
Chapter 4.

PRELIMINARY WATER BALANCE

Groundwater is an important component in the water budget or water balance of a drainage basin. The simplest water balance equation for a watershed for any period would take the form

\[ I - O = \frac{ds}{dt} \]  

(4.1)

which states that inflow I into a basin minus outflow O out of the basin must equal the change in the storage of water \( \frac{ds}{dt} \) within the basin.

The input and output can be broken down into their major components. The main input parameter is precipitation and the important output parameters are evapotranspiration and runoff out of the basin. The runoff has two prominent components, the surface water contributions and the groundwater contributions.

A water balance for the entire Lillooet drainage basin is attempted, to estimate the contribution of groundwater discharge to the total runoff. As a steady-state situation is being assumed, where recharge equals discharge, then the amount of water entering the groundwater system is also being estimated. This information is useful in two ways. Firstly, in Chapter 6, an estimate of groundwater recharge is calculated for the Meager Creek sub-basin. The calculations are made in a different manner than in the water balance of this chapter. If the two estimates of groundwater recharge give
similar results then it can be assumed that they are reasona-
ably correct. Secondly, in Chapter 7, the amount of groundwa-
ter recharge is an important variable as an input parameter 
in the mathematical model. If the value of the groundwater 
recharge is known within a small range, it will narrow the 
possible range of flow values of the other variables that 
influence groundwater.

To develop a water balance for the Lillooet River basin 
the precipitation, temperature, evapotranspiration, and 
runoff of the basin will have to be examined.

Precipitation and Temperature

The first temperature-rainfall stations in the study 
area were installed in the summer of 1980, therefore no data 
of value is yet available. The closest meteorological sta-
tions to Meager Mountain exist at Pemberton Meadows, 50 km to 
the southeast, and Bralorne, 60 km to the northeast
(Fig. 4.1). The mean daily temperatures and mean total pre-
cipitation for these two sites are tabulated in Table 4.1.
The mean total precipitation at Pemberton Meadows is 1024 mm 
and at Bralorne is 732 mm. The mean daily temperature at 
Pemberton Meadows is 7.2°C and at Bralorne is 4.3°C. The 
values are averaged from 1941 to 1970, the current standard 
30 year period recognized by the World Meteorological 
Organization (Environment Canada, 1974).

The elevation of the valley areas around Meager Mountain
Figure 4.1 Location of data gathering stations in the Lillooet River area referred to in the text.
Table 4.1
Mean total precipitation and mean daily temperature at Pemberton Meadows and Bralorne 1941-1970

<table>
<thead>
<tr>
<th>Month</th>
<th>Pemberton Meadows</th>
<th>Bralorne</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(El. 300 m)</td>
<td>(El. 1300 m)</td>
</tr>
<tr>
<td></td>
<td>Mean Daily Temperature (°C)</td>
<td>Mean Total Precipitation (mm)</td>
</tr>
<tr>
<td>Jan</td>
<td>-6.0</td>
<td>168</td>
</tr>
<tr>
<td>Feb</td>
<td>-1.7</td>
<td>85</td>
</tr>
<tr>
<td>Mar</td>
<td>2.6</td>
<td>64</td>
</tr>
<tr>
<td>Apr</td>
<td>8.2</td>
<td>45</td>
</tr>
<tr>
<td>May</td>
<td>13.4</td>
<td>31</td>
</tr>
<tr>
<td>June</td>
<td>16.0</td>
<td>38</td>
</tr>
<tr>
<td>July</td>
<td>18.6</td>
<td>27</td>
</tr>
<tr>
<td>Aug</td>
<td>17.1</td>
<td>28</td>
</tr>
<tr>
<td>Sept</td>
<td>13.4</td>
<td>64</td>
</tr>
<tr>
<td>Oct</td>
<td>7.6</td>
<td>141</td>
</tr>
<tr>
<td>Nov</td>
<td>0.8</td>
<td>161</td>
</tr>
<tr>
<td>Dec</td>
<td>-3.4</td>
<td>179</td>
</tr>
<tr>
<td>TOTAL</td>
<td>7.2</td>
<td>1024</td>
</tr>
</tbody>
</table>

(From Climate of British Columbia, Climatic Normals, 1941-1970)
lie at 760 m, whereas those at Pemberton Meadows and Bralorne lie at 300 m and 1300 m respectively. The Meager Mountain temperature regime in the valleys probably lies between the temperature regime of these two stations.

It has been found that precipitation is extremely variable in the coastal mountains. In general, the amount of precipitation drops off dramatically from west to east and from high elevations to lower. The west-east variation is a result of moisture laden air from the Pacific dropping the majority of its water on the most westerly mountains. The mean total precipitation at Pemberton Meadows is 1024 mm while the mean annual runoff is 1880 mm over the drainage basin. This indicates that a great deal more precipitation falls at higher elevations than in the valley areas. This point will be expanded upon in the runoff section.

Evapotranspiration

No direct measurement of evapotranspiration or evaporation are available for the Meager Mountain region. Bruce and Weisman (1967) collected evaporation pan data, observed solar radiation data and estimated solar radiation data from across Canada. They state that potential evapotranspiration is approximately equal to free-water evaporation from small lakes and reservoirs, therefore their annual evaporation values can be used as evapotranspiration values. Their maps indicate that the Meager Mountain area has a maximum potential rate of
evapotranspiration of approximately 500 mm/year.

This value will be used for the water balance calculations later in this chapter.

Runoff

The only long-term measurement of runoff in the Lillooet River basin is at the Lillooet River site, 1.5 km north of Pemberton (Fig. 4.1) where daily discharge rates have been measured by the Water Survey of Canada since 1914. Figure 4.2 illustrates average runoff, temperature and precipitation of the Pemberton Valley. It is clear, that runoff correlates with temperature and not precipitation. This is due to the large amount of runoff which results from melting snow and ice at higher elevations in the warm summer months. Note, that runoff is given in [L]. The volume [L³] of water discharging out of the basin is measured and divided by the area of the basin [L²] to give a runoff value over the drainage basin of dimension [L].

As mentioned in the precipitation section, the mean annual runoff of the Lillooet River from 1923-1973 was 1880 mm while the average precipitation at the Pemberton Meadows station from 1941-1970 was 1024 mm. This indicates that precipitation at higher elevations is much greater than 1880 mm/year.

Mean annual runoff in the Miller Creek basin, a tribu-
Figure 4.2 Hydrometeorological regime of the Lillooet River and valley bottom (after Teti, 1979). Temperature and precipitation were recorded at Pemberton Meadows (1931-1960). Discharge was recorded about 1.5 km north of Pemberton (1923-1973).
tary of the Lillooet just north of Pemberton is estimated to be 2400 mm ± 300 mm. This is based on summer discharge records and visual estimates of discharge during winter low flow (Slaymaker, 1973, 1974 and 1975, unpublished data). Precipitation in the Meager Mountain area at higher elevations may be 3000 mm/year or possibly higher.

With estimates of precipitation, evapotranspiration and runoff at hand, water balance calculations can now be performed.

**Water Balance**

In the Meager Mountain area, a much more detailed water balance equation than the previously-mentioned Equation 4.1 is needed in order to quantify all the hydrologic components. Assuming the surface-water divides and groundwater divides coincide, there is no change in groundwater storage, and there are no external inflows or outflows of groundwater, then a water budget equation would be written for an annual period as follows,

\[ P = Q_{sp} + Q_{sa} - Q_{sac} + Q_{sbm} + Q_g + E_s + E_w \]

(4.2)

where \( P \) is the average annual precipitation, \( Q_{sp} \) is the surface water component of average annual runoff due to precipitation, \( Q_{sa} \) is the surface water component of average annual runoff due to glacial ablation, \( Q_{sac} \) is the surface water component of average runoff due to glacial accumulation, \( Q_{sbm} \) is the surface water component of average annual runoff due
to glacial basal melting, \( Q_g \) is the groundwater component of average annual runoff, \( E_s \) is the evapotranspiration from the soil-covered portion of the basin, and \( E_w \) is the evaporation from the open water portion of the basin.

Equation 4.2 includes the major components necessary for a rigorous and accurate water balance calculation in the Meager Mountain area. Data is not available for all the variables, but a number of simplifying assumptions can be made to make the problem tractable. In the first place, very little open water exists in the basin, so \( E \) can be considered negligible. In addition, the component \( Q_{sbm} \) represents the continual basal melting of glaciers due to the natural geothermal gradient of the portion of the earth the glacier covers. The governing equation is

\[
\frac{dz}{dt} = \frac{Q_o}{\rho L}
\]

where \( \frac{dz}{dt} \) is the rate of change of thickness of ice with time, \( Q_o \) is the geothermal heat flux, \( \rho \) is the density of ice and \( L \) is the latent heat of melting (personal communication, G. Clark, 1981). Lewis and Souther (1978) discovered, that in the Meager Creek basin the average geothermal heat flux is approximately 10 times the world average. Calculations reported in Appendix II demonstrate that \( Q_{sbm} \) contributes a maximum of 3 to 4% of the total runoff in the Meager Creek basin. The basal melt contribution for the entire Lillooet basin which has an overall heat flux close to the world average is less than 1%, and is therefore assumed to be insignificant.
Measurements of glacier ablation or accumulation have not been undertaken in the study area, however it has been shown by historical, botanical and geological studies that glaciers in the Coast Mountains of British Columbia have been retreating since the second and third decades of this century (Mathews, 1951). This retreat has slowed somewhat since the nineteen fifties. The rate of ablation of a number of glaciers in southwestern British Columbia has been investigated. Mathews (1951) examined a number of glaciers in the Mount Garabaldi area 40 to 80 km north of Vancouver. The most complete data exists for the Helm Glacier, which was found to be lossing 1.8 to 2.1 m/y of water equivalent thickness between 1928 and 1947. The glacier nearest the study area that has been studied is the Place Glacier 20 km northeast of Pemberton. From 1965 to 1974 the glacier lost an average of 0.4 m/y (Mokievsky-Zubok and Stanley, 1976).

The runoff data available for the Lillooet basin (Figure 4.3) is averaged from 1923 to 1968. This time period covers the rapid ablation period early in the century and the slower ablation period in recent time. Therefore, an average ablation rate of 1 m/y has been chosen for the glaciers of the Lillooet basin. Glaciers cover 14% of the surface area of the basin. Consequently, assuming that all the glacier ablation takes place as runoff, the glaciers contribute 0.14 m/y or 140 mm/y over the entire drainage basin.

The above considerations allow Equation 4.2 to be re-
duced to
\[ P = Q_{sp} - Q_{sg} + Q_{g} + E \]  \hspace{1cm} (4.4)
The fact that \( E \) is the only evapotranspiration term allows Equation 4.4 to simplify to

\[ P = Q_{sp} - Q_{sg} + Q_{g} + E \]  \hspace{1cm} (4.5)
where \( E \) is the average annual evapotranspiration.

The average annual runoff, \( Q_{sp} + Q_{sg} + Q_{g} \), over a 50 year period was given earlier as 1880 mm and the evapotranspiration \( E \) was estimated at 500 mm. Adding the above values and subtracting \( Q_{sg} \) gives an average annual precipitation \( P \), in the Lillooet basin, of approximately 2240 mm.

We would also like to gain some idea of the relative values of \( Q_{s} \) and \( Q_{g} \). During the months of January and February the mean annual temperature in the Lillooet basin is below freezing (Table 4.1), so that all runoff presumably comes from groundwater discharge into the streams. Figure 4.3 represents the mean of the mean daily discharge of the Lillooet River from 1923 to 1968. The groundwater component or baseflow of the runoff can be calculated from the graph. The simplest technique of baseflow separation is to draw a horizontal line through the point at which surface runoff begins, point A (Viessman et al., 1977). This method assumes a steady-state groundwater flow system. That is to say, the groundwater flow is constant throughout the year, thereby justifying the horizontal line. The area below the line represents the groundwater component of the average annual
Figure 4.3 Mean of mean daily discharge for Lillooet River (1923-1968) divided into surface water component $Q_{SP} + Q_{SA}$ and groundwater component $Q_g$. 
runoff $Q_g$. The area above the line represents the surface water component of the average annual runoff $Q_{sp} + Q_{sac}$. The area below the horizontal line is 20% of the total area below the curve, therefore the groundwater component $Q_g$ is 20% of the average annual runoff of 1880 mm, or 380 mm.

In conclusion, the preliminary simplified water balance equation for the Lillooet basin is

$$P = Q_{sp} - Q_{sac} + Q_g + E$$

(4.5)

where the average annual precipitation $P$ equals 2240 mm and 61% or 1360 mm of this total is released from the basin by the surface water component of the average annual runoff due to precipitation $Q_{sp}$, 17% or 380 mm by the groundwater component of the average annual runoff $Q_g$, and 22% or 500 mm by the evapotranspiration $E$. The surface water component of the average annual runoff due to glacial ablation $Q_{sac}$ is 140 mm/y.

The concluding equation indicates that 17% of the precipitation falling in the Lillooet River basin enters the groundwater system. In Chapter 6, this value of groundwater recharge will be compared with the value calculated in the Meager Creek sub-basin by a very different method. In Chapter 7, the values are used as input parameters in the mathematical model.

First, we will discuss the fracture survey completed at Meager Mountain in an attempt to calculate the hydraulic conductivity of the fractured basement rock.
Chapter 5.

HYDRAULIC CONDUCTIVITY OF FRACTURED ROCK

The hydraulic conductivity is a very important parameter for determining groundwater flow patterns. As stated earlier, the hydraulic conductivity in the Meager Mountain area is governed mainly by the fracture porosity of the volcanic and basement rocks. Many researchers have developed formulas that relate the fracture porosity $n$ to the hydraulic conductivity $K$. Snow (1968) illustrated that a parallel array of planer joints of aperture $b$, with $N$ joints per unit distance across the rock face has a porosity $n=Nb$ and hydraulic conductivity:

$$K = \left( \frac{\rho g}{\mu} \right) \left( \frac{Nb^3}{12} \right)$$  \hspace{1cm} (5.1)

or

$$k = \left( \frac{Nb^3}{12} \right)$$  \hspace{1cm} (5.2)

where $k$ is the permeability of the rock, $\rho$ is the density of the fluid, $g$ is the acceleration due to gravity and $\mu$ is the dynamic viscosity of the fluid. Therefore, the only field measurements necessary are the aperture and spacing. Note that the hydraulic conductivity varies as the cube of the aperture. It is clear, that inaccurate measurements of fracture apertures will create large errors in the estimated conductivity.

More complex equations have been developed for two and three dimensional anisotropic flow in fractured rock. In addition to aperture and spacing, strike and dip measurements
are also needed. The equations and their development will not be discussed here but the interested reader is referred to Snow (1966,1969), Bianchi and Snow (1969), Rocha and Franciss (1977), and Streltsova (1976).

During the summer of 1980 a fracture survey was undertaken at Meager Mountain to evaluate the hydraulic conductivity of the basement rock. The results of this fracture survey are discussed below. To check the validity of the survey, the results are compared with published data of the fracture permeability of various rock types.

Firstly, however, proper field mapping and data processing techniques for the hydraulic conductivity determination of fractured rocks will be discussed.

Fracture Mapping and Data Processing Methods

The fracture system in a rock mass contains discontinuities of various sizes. The largest fractures are large-scale structural features such as major faults and shear zones. These features are seldom observed in surface bedrock exposures but are easily identified by aerial photography. The most common type of fractures are joints which are discontinuous discrete breaks within the rock mass, commonly occurring in sets reflecting the tectonic history of the rock mass (Raven 1980). Transitional between major faults and joints are fracture zones composed of closely spaced interconnecting breaks. Such zones may have widths of meters to
tens of meters.

A number of methods for mapping fracture systems have been developed. A method in common use is the line sampling technique originally employed for excavation stability analysis by Piteau (1971). The method consists of stretching a measuring tape along an exposed rock face and recording the geological discontinuities that intersect the tape. The measurements and features that can be recorded for each discontinuity include:

1) distance from traverse station to structure
2) rock types
3) hardness
4) type of structures (fault, joint etc)
5) spacing and frequency of joint sets
6) aperture
7) strike
8) dip
9) length of discontinuity
10) infilling
11) water
12) roughness
13) waviness

Piteau and Martin (1977) include genetic descriptions of the above terms to insure uniform recording by different individuals.

Mapping along level lines on near vertical outcrops neg-
lects fractures in the horizontal direction. Fracture information from boreholes can be treated as line samples to complete the three-dimensional assessment of fracture orientation.

The aperture of a discontinuity is one of the most important parameters with respect to hydraulic conductivity determination, and also the most difficult parameter to measure in the field. Snow (1970) discusses a technique that involves photography and fluorescent liquid dye. Through the use of dye, solvents, and a developer, fractures can be highlighted. Close up photos are taken and measurements of the fractures are made with calipers on blown up pictures projected on a screen. There are a few drawbacks to this method but apertures as small as 130 microns can be measured with a relative error of 3 percent.

The amount of data gathered during a fracture survey can be large and difficult to manage. Several computer based systems for the storage, retrieval, analysis and display of structural discontinuity data have been developed. Cruden, Ramsden and Herget (1977) developed a computer based package called DISCODAT. It is described in a part of the Pit Slope Manual series published by Energy, Mines and Resources Canada. Both surface and borehole fracture data may be analyzed with the Discodat programs. Orientation diagrams such as stereo nets, and histograms of relative and cumulative frequencies of the angle of strike are produced by the
Discodat system, saving countless hours of tedious manual plotting. The line sampling technique for gathering data and the Discodat system work well together to produce the necessary data for fractured-rock-hydrology studies.

**Meager Mountain Fracture Survey**

During the summer of 1980 the author and the employees of Nevin, Sadlier-Brown and Goodbrande Ltd. performed a joint orientation survey on the basement granodiorite in the South Reservoir area. The purpose of the study for Nevin, Sadlier-Brown and Goodbrande Ltd. was to gain structural data for geologic interpretation and for the author to collect data for hydraulic conductivity calculations.

**Results**

Twenty-five locations were examined. Twenty-four sites were in basement rock and one in volcanic rock. Figure 5.1 illustrates the site locations. The line sampling technique described earlier was not utilized in the survey because the author was not aware of the technique at the time of the survey. Rather, traverses were made along outcrops and joints were measured in a random fashion. The spacing between adjacent joints of the same set and the aperture of each fracture were estimated rather than measured directly.

Lower hemisphere, equal-area, contoured-pole plots of the twenty-five sites can be found in Appendix II along with
Figure 5.1 Location of fracture survey sites.
the coordinates and elevation of each site. Close examination of the contoured-pole plots revealed four persistent joint sets that have been highlighted for easy identification. Table 5.1 exhibits the variation in strike and dip of individual joint sets at locations where joint sets were evident while Table 5.2 illustrates the average orientation of the joint sets at each measured site. Note that joint sets 1 and 2 exist at almost every site while joint sets 3 and 4 occur only at a few sites, indicating that joint sets 1 and 2 are the dominant sets in the rock. It is also clear that the joint sets are also steeply dipping.

The variation in spacing and aperture of joint sets can be found in Table 5.3, and the average variations of spacing and aperture of the four joint sets are in Table 5.4. It is evident that there is a wide range of spacings from 0.05 to 5.0 m and of apertures from tight to 20 mm. Therefore, the hydraulic conductivity which is calculated from these values will also have a wide range. The list is incomplete due to the difficulty of seeing the individual joint sets on the rock faces at a number of locations. Consequently, the joint spacing parameter could not be estimated. Also, the author's neglect to stress the importance of the spacing and separation observations to others taking measurements, led to missing values at a number of sites.
TABLE 5.1
Orientation Variations of Joint Sets

<table>
<thead>
<tr>
<th>Site</th>
<th>Joint Set 1</th>
<th>Joint Set 2</th>
<th>Joint Set 3</th>
<th>Joint Set 4</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Strike</td>
<td>Dip</td>
<td>Strike</td>
<td>Dip</td>
</tr>
<tr>
<td>1</td>
<td>350-35</td>
<td>65W-88E</td>
<td>106-130</td>
<td>34 -78NE</td>
</tr>
<tr>
<td>2</td>
<td>45-75</td>
<td>70N-75S</td>
<td>105-140</td>
<td>35 75SW</td>
</tr>
<tr>
<td>3</td>
<td>32-68</td>
<td>45NW-75SE</td>
<td>140-160</td>
<td>60 -90SE</td>
</tr>
<tr>
<td>4</td>
<td>17-45</td>
<td>75SE-85SW</td>
<td>105-140</td>
<td>35 75SW</td>
</tr>
<tr>
<td>5</td>
<td>4-25</td>
<td>70W-80E</td>
<td>75-105</td>
<td>35 -65N</td>
</tr>
<tr>
<td>6</td>
<td>105-140</td>
<td>58 -72SW</td>
<td>70-90S</td>
<td>65 -88S</td>
</tr>
<tr>
<td>7</td>
<td>148-170</td>
<td>70W-75E</td>
<td>75-105</td>
<td>35 -65N</td>
</tr>
<tr>
<td>8</td>
<td>130-170</td>
<td>65SW-80NE</td>
<td>70-90S</td>
<td>65 -88S</td>
</tr>
<tr>
<td>9</td>
<td>154-166</td>
<td>60SW-80NE</td>
<td>68-88</td>
<td>60 -88SE</td>
</tr>
<tr>
<td>10</td>
<td>148-170</td>
<td>70NW-80SW</td>
<td>110-158</td>
<td>40 -70SW</td>
</tr>
<tr>
<td>11</td>
<td>136-162</td>
<td>70NW-80SW</td>
<td>108-140</td>
<td>44 -75NE</td>
</tr>
<tr>
<td>12</td>
<td>150-165</td>
<td>62 -90NE</td>
<td>90-116</td>
<td>40 -80NE</td>
</tr>
<tr>
<td>13</td>
<td>78SE-84NW</td>
<td>70NW-80SW</td>
<td>98-128</td>
<td>75NW-70SW</td>
</tr>
<tr>
<td>14</td>
<td>42 -64NW</td>
<td>60SW-78NE</td>
<td>80-122</td>
<td>50 -90NE</td>
</tr>
<tr>
<td>15</td>
<td>52 -78E</td>
<td>75NE-70SW</td>
<td>72NW-80NW</td>
<td>50 -90NE</td>
</tr>
<tr>
<td>16</td>
<td>70SE-80NW</td>
<td>40 -80SE</td>
<td>72-118</td>
<td>56 -80NE</td>
</tr>
<tr>
<td>17</td>
<td>70NW-68SE</td>
<td>70SE-72SE</td>
<td>70NW-62SE</td>
<td>68-82</td>
</tr>
<tr>
<td>18</td>
<td>70NW-68SE</td>
<td>70NW-62SE</td>
<td>70NW-62SE</td>
<td>68-82</td>
</tr>
<tr>
<td>19</td>
<td>70NW-68SE</td>
<td>70NW-62SE</td>
<td>70NW-62SE</td>
<td>68-82</td>
</tr>
<tr>
<td>20</td>
<td>148-176</td>
<td>70SW-70NE</td>
<td>145-174</td>
<td>50NE-80SW</td>
</tr>
<tr>
<td>21</td>
<td>154-166</td>
<td>60SW-78NE</td>
<td>125-162</td>
<td>45 -72SW</td>
</tr>
<tr>
<td>22</td>
<td>144-170</td>
<td>68NE-80SW</td>
<td>88-120</td>
<td>56 -80NE</td>
</tr>
<tr>
<td>23</td>
<td>144-170</td>
<td>68NE-80SW</td>
<td>88-120</td>
<td>56 -80NE</td>
</tr>
<tr>
<td>24</td>
<td>144-170</td>
<td>68NE-80SW</td>
<td>72-118</td>
<td>20 -40N</td>
</tr>
<tr>
<td>25</td>
<td>144-170</td>
<td>68NE-80SW</td>
<td>68-96</td>
<td>64 -90S</td>
</tr>
<tr>
<td>Site</td>
<td>Joint Set 1</td>
<td>Joint Set 2</td>
<td>Joint Set 3</td>
<td>Joint Set 4</td>
</tr>
<tr>
<td>------</td>
<td>-------------</td>
<td>-------------</td>
<td>-------------</td>
<td>-------------</td>
</tr>
<tr>
<td></td>
<td>Strike</td>
<td>Dip</td>
<td>Strike</td>
<td>Dip</td>
</tr>
<tr>
<td>1</td>
<td>12</td>
<td>82 NW</td>
<td>118</td>
<td>56 NE</td>
</tr>
<tr>
<td>2</td>
<td>60</td>
<td>88 NW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>50</td>
<td>75 NW</td>
<td>159</td>
<td>87 SW</td>
</tr>
<tr>
<td>4</td>
<td>31</td>
<td>85 SE</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>127</td>
<td>52 SW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>15</td>
<td>85 SE</td>
<td>135</td>
<td>65 SW</td>
</tr>
<tr>
<td>7</td>
<td>119</td>
<td>55 SW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>150</td>
<td>82 SW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>33</td>
<td>88 NW</td>
<td>160</td>
<td>80 SW</td>
</tr>
<tr>
<td>10</td>
<td>32</td>
<td>79 NW</td>
<td>159</td>
<td>85 NE</td>
</tr>
<tr>
<td>11</td>
<td></td>
<td></td>
<td>158</td>
<td>76 NE</td>
</tr>
<tr>
<td>12</td>
<td>15</td>
<td>87 SE</td>
<td>149</td>
<td>85 NE</td>
</tr>
<tr>
<td>13</td>
<td>54</td>
<td>60 SE</td>
<td></td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>52</td>
<td>53 NW</td>
<td>129</td>
<td>86 SW</td>
</tr>
<tr>
<td>15</td>
<td></td>
<td></td>
<td>113</td>
<td>88 SW</td>
</tr>
<tr>
<td>16</td>
<td>24</td>
<td>85 SE</td>
<td>106</td>
<td>70 NE</td>
</tr>
<tr>
<td>17</td>
<td>14</td>
<td>86 NW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>18</td>
<td>89 SE</td>
<td>134</td>
<td>55 SW</td>
</tr>
<tr>
<td>19</td>
<td>29</td>
<td>88 SE</td>
<td>124</td>
<td>59 NE</td>
</tr>
<tr>
<td>20</td>
<td></td>
<td></td>
<td>162</td>
<td>90</td>
</tr>
<tr>
<td>21</td>
<td>6</td>
<td>86 SE</td>
<td></td>
<td></td>
</tr>
<tr>
<td>22</td>
<td></td>
<td></td>
<td>159</td>
<td>75 NE</td>
</tr>
<tr>
<td>23</td>
<td>54</td>
<td>86 SE</td>
<td>144</td>
<td>59 SW</td>
</tr>
<tr>
<td>24</td>
<td></td>
<td></td>
<td>157</td>
<td>84 NE</td>
</tr>
<tr>
<td>25</td>
<td>32</td>
<td>80 NW</td>
<td>140</td>
<td>85 SW</td>
</tr>
</tbody>
</table>
**Table 5.3**

Joint Spacing and Aperture Variations

<table>
<thead>
<tr>
<th>Site</th>
<th>Joint Set</th>
<th>Spacing (m)</th>
<th>Aperture (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>0.2-1.0</td>
<td>Tight -3</td>
</tr>
<tr>
<td>2</td>
<td>1</td>
<td>0.1-0.5</td>
<td>1-5</td>
</tr>
<tr>
<td>3</td>
<td>1</td>
<td>0.5-1.0</td>
<td>Tight -5</td>
</tr>
<tr>
<td>3</td>
<td>2</td>
<td>0.1-1.5</td>
<td>5-2</td>
</tr>
<tr>
<td>4</td>
<td>1, 3 &amp; 4</td>
<td>0.4-2.0</td>
<td>Tight</td>
</tr>
<tr>
<td>8</td>
<td>2</td>
<td>1.0-5.0</td>
<td>Tight -5</td>
</tr>
<tr>
<td>9</td>
<td>2</td>
<td>0.1-0.5</td>
<td>Tight -10</td>
</tr>
<tr>
<td>9</td>
<td>4</td>
<td>0.05-0.25</td>
<td>Tight -5</td>
</tr>
<tr>
<td>10</td>
<td>1</td>
<td>0.05-1.0</td>
<td>5-20</td>
</tr>
<tr>
<td>10</td>
<td>2</td>
<td>0.05-0.8</td>
<td>Tight -5</td>
</tr>
<tr>
<td>11</td>
<td>2</td>
<td>1-1.5</td>
<td>2-20</td>
</tr>
<tr>
<td>14</td>
<td>2</td>
<td>0.2-1.0</td>
<td>Tight -5</td>
</tr>
<tr>
<td>14</td>
<td>1</td>
<td>0.3-0.5</td>
<td>Tight -3</td>
</tr>
<tr>
<td>15-18</td>
<td>1 &amp; 2</td>
<td>0.1-1.0</td>
<td>1-10</td>
</tr>
</tbody>
</table>

**Table 5.4**

Average Spacing and Aperture

<table>
<thead>
<tr>
<th>Set</th>
<th>Spacing (m)</th>
<th>Aperture (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.1 -1.0</td>
<td>Tight -20</td>
</tr>
<tr>
<td>2</td>
<td>0.1 -1.0</td>
<td>Tight -20</td>
</tr>
<tr>
<td>3</td>
<td>0.4 -2.0</td>
<td>Tight</td>
</tr>
<tr>
<td>4</td>
<td>0.05-2.0</td>
<td>Tight -5</td>
</tr>
</tbody>
</table>
Discussion

The methodology used for the Meager Mountain fracture survey supplied inadequate data for the calculation of the rock mass hydraulic conductivity. The problems with the survey are listed below together with suggestions for possible future remedies.

1) The problem of the missing data for spacing and aperture measurements could be rectified by coordinating and standardizing survey practice before beginning the work.

2) The surveys were taken on near-vertical outcrops, consequently the measured fractures represent only two dimensions. The gathering of joint orientations in inclined bore holes or near horizontal rock faces would correct this deficiency.

3) The random measuring of the joint systems can bias the results, whereas the line sampling technique eliminates any bias.

4) The major problem with the measurements taken at Meager Mountain lies in the fact that the apertures of the fractures were estimated rather than measured directly. The conductivity of the rock mass varies as the aperture cubed, therefore inaccuracies in aperture measurement are multiplied when calculating the hydraulic conductivity or flux through the rock.

Simple hydraulic conductivity calculations of the rock mass were attempted using Equation 5.1. However, due to the
variability of aperture estimates at any single site, the hydraulic conductivity estimates (10^4 to 10^6 m/s) extended over many orders of magnitude and such estimates are felt to have little value. Even if more accurate apertures had been measured, it is questionable whether the measurement of fractures on surface exposures is adequate for computing the conductivity of rock at depth. A near-surface or surface fracture is widened during its history, by a decrease of the stress field by erosional unloading and by surface processes such as weathering and creep (Bianchi and Snow, 1969). Snow (1968) discovered from test holes in granitic, metamorphic and volcanic rock, that within the first 150 m of depth, the fractured rock porosity decreases an order of magnitude every 60 m. Therefore, surface measurements of apertures will give an unreasonably high conductivity for the rock mass. However, fracture data can be used to illustrate general trends of the conductivity in a rock mass, such as the direction of anisotropy. Also, areas of relatively high fracture density can be delineated to possibly reveal zones of higher hydraulic conductivity at depth.

In the later stages of exploration it may be possible to measure apertures in-situ in boreholes with a down-hole camera or periscope, and of course, once a hole is drilled, direct tests for hydraulic conductivity such as pump tests and packer tests can be performed.

As, the fracture survey at Meager Mountain was unsucces-
sful, let us now look at published fracture permeabilities of various rock types, to obtain a range of hydraulic conductivity values that one might expect in the study area.

**Published Fracture Permeabilities of Various Rock Types**

Reported measurements of hydraulic conductivities in fractured rock have been very scarce until recently, when research for feasible nuclear waste repositories was undertaken in the 1970's. Work performed for the Atomic Energy of Canada Limited (AECL) and the U.S. Department of Energy has provided most of the published values.

Raven and Gale (1977), in a report for the Geological Survey of Canada and AECL, examined the surface and subsurface structures and groundwater conditions at 25 underground mines, located on the Canadian Shield. It was found that seepage was restricted to the upper 300 m if no major structures such as faults or contacts were present. They also found that the number and magnitude of these structural seepage zones were greatly reduced below 300 to 350 m. Structures such as joint sets continued to depths of up to 1000 m but the permeabilities of the rock near surface decreased logarithmically with depth. In terms of hydraulic conductivity, faults provide the highest values, followed by shear zones, dikes, sills and intrusive geologic contacts.

Of the 25 mine sites examined, the most extensive program of down hole hydraulic conductivity testing was carried
out at the Gull Island project site in Labrador. The holes were drilled in foliated granodiorite, hornblende biotite gneiss, and amphibolite. Near surface, the hydraulic conductivity varied from $10^{-5}$ to $10^{-4}$ m/s and decreased to $10^{-6}$ to $10^{-8}$ m/s at depths of 60 to 75 m. From 60 to 150 m the hydraulic conductivity showed no significant variation.

Burgess (1979), in another AECL report, discusses constant inflow tests in 8 boreholes completed for the Swedish Radioactive Waste Program. The test holes were sunk over 500 m into a granitic pluton in the Precambrian basement. Hydraulic conductivity values in the upper 100 m varied from $10^{-5}$ to $10^{-7}$ m/s, decreasing to $10^{-8}$ to $10^{-10}$ m/s from 200 to 500 m.

Raven (1979) collected hydrogeologic data for the Nuclear Fuel Waste Management Program at the Chalk River (Ontario) Nuclear Laboratories field research site in 1977 and 1978. Various shut-in pressure and injection tests were accomplished in 5 test holes drilled in Precambrian monzonite gneiss and metagabbro. Hydraulic conductivity values of $4 \times 10^{-9}$ to $4 \times 10^{-11}$ were obtained at depths between 40 and 72 m.

Davison et al. (1979) reported on various down hole tests performed in a granitic pluton at the Whiteshell Nuclear Research Establishment in Manitoba. The tests were completed in two drill holes each 150 m deep. Above 25 m hydraulic conductivity values of $5 \times 10^{-8}$ to $5 \times 10^{-9}$ m/s were
obtained. Below 25 m the hydraulic conductivity ranged from $5 \times 10^{-10}$ to $5 \times 10^{-11}$ m/s.

Over the past 10 to 15 years, extensive drilling and hydrogeologic testing has been undertaken at the Hanford, Basalt Waste Isolation Project in Washington State. The geology consists basically of a number of thick basaltic flows with minor brecciated or weathered horizons and sedimentary interbeds. The maximum variations in the hydraulic conductivity are from $5 \times 10^{-1}$ m/s in very sandy weathered layers to $5 \times 10^{-13}$ m/s in very dense columnar basalt zones. The majority of the weathered, brecciated zones have hydraulic conductivity values ranging from $10^{-3}$ to $10^{-5}$ m/s. Most of the low density basalts range from $10^{-4}$ to $10^{-10}$ m/s and the high density basalt flows vary from $10^{-9}$ to $10^{-16}$ m/s.

Davis (1969) reported in his paper on the porosity and permeability of various materials including plutonic, volcanic, metamorphic and sedimentary rocks and unconsolidated deposits. His findings on the hydraulic conductivity of a number of rock types are summarized in Table 5.5 along with the other hydraulic conductivities discussed above.

The preceding values imply that the hydraulic conductivity of most fractured rock lies in the $10^{-7}$ to $10^{-11}$ m/s range. It should be noted that, the majority of the values given in the various reports were taken in reasonably undisturbed rock. However, the basement rocks of the Meager Mountain area have been highly disturbed. Therefore, it
Table 5.5
Summary of Measured Hydraulic Conductivity Values for Various Rock Types

<table>
<thead>
<tr>
<th>Reference</th>
<th>Location</th>
<th>Rock Type</th>
<th>Depth (m)</th>
<th>Conductivity (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Raven and Gale 1977</td>
<td>Gull Island</td>
<td>Foliated granodiorite</td>
<td>20</td>
<td>$10^{-5}$ to $10^{-6}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>and biotite gneiss</td>
<td>60-75</td>
<td>$10^{-6}$ to $10^{-8}$</td>
</tr>
<tr>
<td>Burgess 1979</td>
<td>Sweden</td>
<td>Granite pluton</td>
<td>0-100</td>
<td>$10^{-5}$ to $10^{-7}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>200-500</td>
<td>$10^{-8}$ to $10^{-10}$</td>
</tr>
<tr>
<td>Raven 1979</td>
<td>Chalk River Ontario</td>
<td>Monzonite gneiss</td>
<td>-</td>
<td>$4 \times 10^{-9}$ to $4 \times 10^{-11}$</td>
</tr>
<tr>
<td>Davison, Grisak and Williams 1979</td>
<td>Whiteshell Manitoba</td>
<td>Granite pluton</td>
<td>25</td>
<td>$5 \times 10^{-8}$ to $5 \times 10^{-9}$</td>
</tr>
<tr>
<td>Rockwell International 1979</td>
<td>Hanford Washington</td>
<td>Basalt</td>
<td>150</td>
<td>$5 \times 10^{-10}$ to $5 \times 10^{-11}$</td>
</tr>
<tr>
<td>Davis 1969</td>
<td></td>
<td>Dense basalt</td>
<td>-</td>
<td>$4 \times 10^{-10}$ to $4 \times 10^{-13}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pumiceous tuff</td>
<td>-</td>
<td>$10^{-10}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Metasediments</td>
<td>-</td>
<td>$10^{-7}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Greywacke</td>
<td>-</td>
<td>$3 \times 10^{-7}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-</td>
<td>$4 \times 10^{-7}$</td>
</tr>
</tbody>
</table>
would be expected that the hydraulic conductivity values of
of the basement rocks are at the high end of the range
stated. Values of $10^{-7}$ to $10^{-8}$ m/s could be expected.

In the hydrogeology chapter and mathematical modelling
chapter to follow, it will be shown that the hydraulic condu-
ctivity of the basement rock apparently falls into this $10^{-7}$
to $10^{-8}$ m/s range.
Chapter 6.

HYDROGEOLOGY

Hydrogeology can be defined as the science that deals with the occurrence, distribution and movement of groundwater. Three important aspects of the hydrogeology of the Meager Mountain area will be addressed. These are the water table configuration, the hydraulic conductivity of the various geological materials, and an estimate of the amount of groundwater recharge. The following interpretation of the hydrogeology is based on the local and regional geology, the location and nature of the hot and cold springs, published hydraulic conductivity values of fractured rock, the water balance calculations, and general field observations.

Figure 6.1 is a cross section through Meager Mountain showing schematically the general flow of groundwater in the system. Line CD represents the location of the groundwater divide in the mountain. All water entering the groundwater zone south of the line will discharge in the Meager Creek valley while all the water entering the system north of the line will discharge into the Lillooet River valley. The water of the Meager Creek Hotsprings has a different flow path than the water of the Pebble Creek Hotsprings, therefore dissimilar water geochemistries found by Hammerstrom and Brown (1977) are not surprising.

The Meager Creek side of the mountain (section CDEF) has greater data availability and more geothermal promise than
Figure 6.1 General groundwater flow in Meager Mountain.
the Lillooet River side, so it will be used for further discussion in this chapter and the following chapter on modelling.

**Water Table Configuration**

The water table configuration is important because it determines the gradient or driving force for the groundwater flow at Meager Mountain. Recall from Chapter 3 that the gradient in Darcy's Law is \( \frac{dh}{dl} \) where \( dh \) is the change in head over a change in distance \( dl \). Consider the two water-table configurations in Figure 6.2. Points, A, C, D and F are on the water table so, by definition, the head at these points is equal to the elevation of the points. The head drop from A to C is much greater than from D to F and yet the change in \( l \) is the same in both cases. Therefore, the gradient is larger when the water table is higher. A higher gradient causes more water to move through the system per unit time, leading to a larger volume of water to discharge in the valley area.

Little is known about the water table configuration in mountainous regions. It is not known whether it sits high in most mountains or is relatively flat. The configuration depends on the hydraulic conductivity of the rock and the amount of recharge available. In areas of rolling hills it has been observed that the water table tends to be a subdued replica of the topography (Fig. 4.6).

A highly elevated water table would lead to the develop-
Figure 6.2 Water table elevation and seepage face development.
ment of seepage faces on the side of the mountain [Fig. 6.2(a)] where the water table intersects the surface. These zones would be the site of many springs. In the field, very few springs were observed and no major seepage areas of large extent occur. One is therefore led to believe that the discharge area for the groundwater flow system is confined to the sections of the valley overlain by unconsolidated Quaternary deposits. It seems likely that the water table is not situated at a high elevation in the mountain but rather occupies a more intermediate position [Fig. 6.2(b)]. The physically possible range of water table configurations will be calculated in the next chapter.

Hydraulic Conductivities of the Geologic Materials

The other parameter governing the groundwater flow, besides the water-table configuration is the hydraulic conductivity \( K \).

To evaluate the geologic materials hydrogeologically they must be divided into groups with similar hydraulic conductivity characteristics. Maxey (1964) defines hydrostratigraphic units as "bodies of rock with considerable lateral extent that compose a geologic framework for a reasonably distinct hydrologic system." As mentioned earlier in Chapter 2, the fracture hydraulic conductivity that occurs across all the volcanic layers has greater control over the groundwater flow than the intergranular hydraulic conductivity dif-
ferences between layers. Consequently, the entire volcanic pile can be considered as one unit with respect to groundwater flow. The basement and the unconsolidated deposits represent two other units that would have characteristic hydraulic conductivities. Therefore, the geological formations in the Meager Mountain area can be divided into three hydrostratigraphic unit; the volcanics, the basement, and the unconsolidated Quaternary deposits. No direct measurements of hydraulic conductivity have been made but a range can be invoked for each hydrostratigraphic unit.

The unconsolidated deposits exist mainly as valley fill in the Meager Creek and Lillooet River valleys. Drilling has revealed that the sediments are mainly sands and boulders with thin layers of clay. Freeze and Cherry (1979) estimate that these deposits should have hydraulic conductivities in the range $10^{-5}$ to $10^{-2}$ m/s.

The hydraulic conductivity of the granodiorite and monzonite basement is due to fracture permeability. Published data discussed previously suggest the highly disturbed basement would have a conductivity from $10^{-7}$ to $10^{-8}$ m/s. The fracture survey revealed that the two dominant fracture sets have a near vertical dip, causing the rock to have a greater conductivity in the vertical direction. If we denote the horizontal hydraulic conductivity by $K_x$ and vertical hydraulic conductivity by $K_z$, then the ratio of $K_x:K_z$ may be as high as 1:5. Accordingly, the vertical hydraulic conduc-
ty in the basement rock probably varies from $10^{-7}$ to $10^{-8}$ m/s while the horizontal hydraulic conductivity probably varies from $2 \times 10^{-8}$ to $2 \times 10^{-9}$ m/s.

The volcanic rocks tend to be a multilayered media of breccia and ash layers interbedded with flow layers. As discussed in Chapter 4 the layering gives the volcanics an overall anisotropy with the conductivity being greatest in the horizontal direction. Consider, the typical assemblage in Figure 6.3 of a basal breccia or tuff overlain by a flow with a thin weathered horizon on the top of the flow. Typical hydraulic conductivity values of the different layers, taken from Table 5.5 include $10^{-7}$ m/s for the breccia, $10^{-8}$ m/s for the flow and $10^{-5}$ m/s for the weathered zone. By invoking Equations 3.9 and 3.10, the calculated equivalent vertical hydraulic conductivity $K_z$ for the assemblage is $1.9 \times 10^{-8}$ m/s and the calculated equivalent horizontal hydraulic conductivity $K_x$ is $3 \times 10^{-7}$ m/s. The anisotropic ratio $K_x : K_z$ is therefore, approximately 16:1. As stated earlier, there is extensive vertical to near vertical fracturing in the volcanics that will reduce the anisotropy possibly as low as 1:1 or isotropic. Anisotropic ratios of greater than 5:1 would not be expected.

The vertical conductivity $K_z$ of the volcanics must be very similar to that of the underlying basement rock. If it were not, one would expect extensive spring lines at their contact; in the field, very few are in evidence. Consequent-
Figure 6.3 Equivalent hydraulic conductivity in layered volcanics.

Figure 6.4 Cross-section of Meager Creek at stage location.
ly, the vertical hydraulic conductivity in the volcanics presumably ranges from $10^{-7}$ to $10^{-8}$ m/s while the horizontal hydraulic conductivity is similar or up to a maximum of 5 times greater.

**Estimates of Groundwater Recharge**

In the Meager Creek basin the only hydrological measurements available in 1980 were river stages recorded at the Meager Creek Hotsprings bridge in 1979. These measurements are reported in Appendix III. A river stage is simply the level of the stream surface. It can be turned into a discharge measurement if the stream bottom configuration is known (this allows the calculation of a cross-section area) and if the stream velocity is known.

On July 4, 1980 a cross section at the river stage station was taken along with a river level reading (Fig. 6.4). Any stage reading taken previously can then be converted to a cross-sectional area. The velocity of the stream, at any stage, can be calculated using the Manning equation. This allows discharge values to be computed for any stage measurement. For an explanation of the Manning equation see Appendix IV.

In Chapter 3 it was noted that during the months of January and February it can be expected that the river discharge will be entirely supplied by groundwater flow. If the discharge at the Meager Creek station during January and
February could be calculated, then the amount of groundwater discharge and recharge taking place in the basin could be estimated. Recall from Chapter 3 that, assuming steady state groundwater flow, the recharge to the system must equal the discharge.

Stage readings were not taken in January or February but one reading taken on December 10, 1979 which calculates to a discharge of 3.7 m$^3$/s. Comparing the discharge at the Lillooet River station north of Pemberton on December 10, 1979 with other December discharges over the past 3 years (Appendix V), it appears to be an average day. We can therefore assume that the measured discharge at the Meager station on December 10 is also a reasonable representation for that time of year.

Note, that two horizontal lines are drawn across Figure 4.3, the graph of the average daily discharge of the Lillooet River. The bottom dashed line, discussed earlier, is the base flow separation line or the average January-February discharge. The upper dashed-dotted lie represents the mean December discharge. It is clear, that the January-February discharge is approximately 70% that of December. Also recall, from Appendix I, that the calculated basal glacier melt flux in the Meager Creek basin supplies 0.09 m$^3$/s or 4% of the runoff. The known December discharge rate must therefore be reduced by 30% to represent January-February discharge and another 4% to represent the groundwater discharge component.
of the runoff. Consequently, the groundwater discharge is 2.5 m$^3$/s or 7.9 m$^3$/y. The measured area of the basin supplying the stage site is 2.18x10$^4$ m$^2$, therefore, the portion of the annual precipitation over the drainage basin that enters the groundwater system is 0.36 m/y or 14.5%; assuming an annual precipitation of 2.5 m/y. This percentage correlates well with the groundwater component of the total annual precipitation calculated from the water balance for the whole Lillooet River (16.5%) in Chapter 4. We can accordingly assume the values of discharge calculated are reasonably correct.

Figure 6.5 schematically illustrates the conclusions on the hydrogeology of the south side of Meager Mountain. Isotropic to horizontally anistropic volcanics overlie a vertically anistropic granodiorite basement. The vertical hydraulic conductivity of the granodiorite and the volcanics is similar. Very permeable, isotropic unconsolidated deposits fill the valley bottom where a groundwater discharge of 0.36 m/y over the drainage basin takes place. The water table has an intermediate position in the mountain complex. These qualitative interpretations of the water table configuration, conductivity ranges, and flux through the system can now be used as input into a mathematical model to quantify our understanding of the hydrogeology.
Figure 6.5 Summary of the hydrogeology on the south of Meager Creek.
Chapter 7.
GROUNDWATER MODELLING

Computer-based numerical methods are one of the major tools used for solving large-scale groundwater forecasting problems (Bear 1979). With the recent advance of computer technology, much effort has been devoted to the development of techniques for numerical solution of the partial differential equations that govern the flow of water in various geological environments. The end product of this research has been a number of computer programs which in most cases are readily available to any user. With modifications a hydrogeologist can usually make an available program applicable to her or his specific problem.

Freeze and Cherry (1979) describe mathematical modelling as a four-stage process involving 1) an examination of the physical problem, 2) replacement of the physical problem by an equivalent mathematical problem, 3) solution of the mathematical problem with accepted mathematical techniques, and 4) interpretation of the mathematical results in terms of the physical problem.

In this study groundwater flow is simulated using a deterministic mathematical model. It is based on physical laws, not on statistical or empirical relationships. The equation of flow for this work is solved for steady-state flow.
Mathematical models of groundwater flow take the form of boundary value problems. To fully define a boundary value problem for steady-state subsurface flow, one needs to know 1) the size and shape of the region of flow, 2) the equation of flow within the region, 3) the boundary conditions around the boundaries of the region, 4) the spatial distribution of hydraulic conductivity values in the region, and 5) a mathematical method of solution (Freeze and Cherry, 1979). All of these aspects will be discussed subsequently in the simulation strategy section. Then, the mathematical modelling results and the significance of the results will be discussed. Firstly, however, an examination of the computer programs used and their capabilities is in order.

The Computer Program

The author has modified and used two different programs for the analysis of groundwater flow in Meager Mountain. The first was written by R.A. Freeze and is known as FEPS; the second was written by S.P. Neuman of the University of Arizona and is known as FREESURF1. A modified version of FEPS prepared by the author has been renamed FOPS.

The two programs have the following similar capabilities. 1) The region of flow is two-dimensional but can have any size or shape; 2) the hydraulic conductivity distribution can take on any desired configuration or range of values; 3) the hydraulic conductivity can take on any direction and
degree of anisotropy; 4) the boundaries can be given any one of five possible boundary conditions. These boundary conditions are, a) impermeable, b) constant specified hydraulic head, c) constant flux in or out of the system, d) free surface (water table) on the upper boundary, and e) seepage face on the upper boundary. In addition to these characteristics, FREESURF1 can handle, 1) three dimensional flow if flow retains an axial symmetry about the vertical coordinate, and 2) interior seepage faces where the free surface becomes discontinuous across an interface between two different materials, as in an earth dam. For background material and examples of the kinds of problems that can be solved with FREESURF1 the reader is directed to Neuman and Witherspoon (1970).

The two most common numerical techniques used to solve boundary value problems in hydrogeology are the finite-difference method and the finite-element method. While different in many ways both techniques operate to solve the groundwater flow equation directly. The finite-element method is utilized by both FOPS and the FREESURF1 programs.

The output from the mathematical models is a field of hydraulic head values at the nodal points of a grid superimposed over the flow field. The output can be contoured and used to construct flow nets. By means of Darcy's law, one can calculate the amount of inflow and outflow along the boundaries of the flow net, or the velocity of the flow at any interior point. By means of Equation 3.6, one can calcu-
late the fluid pressures at any point in the system.

The FOPS model was used for an initial sensitivity analysis. The free surface was set as a constant head boundary, and the hydraulic conductivity field, the depth-of-flow field and the elevation of the water table were varied to observe their effects on the total discharge leaving the system. The FREESURF1 model was then utilized to simulate various possible hydrogeologic environments at Meager Mountain and their effect on the groundwater regime. The recharge rate into the system was varied for each hydrogeologic environment to examine the water table configuration and amount of discharge it produced.

Before discussing the results of the modelling, the boundary value problems we wish to solve using the models, must be defined.

Simulation Strategy

Region of Flow

The region of flow for the Meager Mountain groundwater simulations is a two-dimensional vertical cross-section as illustrated in Figure 7.1. The section was taken along the line A-B in Figure 7.2 through drill holes M5-78D and M7-79D. The lower end of the section is the centre of Meager Creek while the upper end is at the highest point of Capricorn Mountain. The upper boundary is the ground surface and the
Figure 7.1 Region of flow for mathematical modelling.

Figure 7.2 Line of section taken for cross-section in Figure 7.1.
lower boundary for the FREESURF1 simulations was chosen arbitrarily at -2000 m elevation. The lower boundary for the FOPS simulations varied from -305 m to -915 m for the sensitivity analysis. The cross-section is considered to have a unit thickness perpendicular to the page.

Equation of Flow

The equation for saturated, two-dimensional, steady-state flow through heterogeneous, anisotropic material is:

\[
\frac{\partial}{\partial x} \left[ K_x(x,z) \frac{\partial h}{\partial x} \right] + \frac{\partial}{\partial z} \left[ K_z(x,z) \frac{\partial h}{\partial z} \right] = 0
\]  

(7.1)

where \( x \) and \( z \) are the horizontal and vertical coordinates, \( K_x \) and \( K_z \) are the horizontal and vertical components of the anisotropic hydraulic conductivity tensor and \( h \) is the hydraulic head. Assigning \( K_x=K_x(x,z) \) and \( K_z=K_z(x,z) \), identifies the conductivity distribution as heterogeneous.

The solution of Equation 7.1 is the hydraulic head field \( h(x,z) \) in the cross-section.

Boundary Conditions

The boundaries BC, CD and DA of Figure 7.1 are impermeable while AEFB is the water table. In mathematical terms

\[
\frac{\partial h}{\partial x} = 0 \text{ on BC and AD}
\]  

(7.2)

and

\[
\frac{\partial h}{\partial z} = 0 \text{ on CD}
\]  

(7.3)
On AE, the hydraulic head is equal to the elevation of Meager Creek because the water table is very near the surface in the valley fill deposits, therefore,

$$H = Z_r \text{ on AE}$$  \hspace{1cm} (7.4)

where $Z_r$ is the elevation of Meager Creek.

The water table EFB is treated in two ways. When using the FOPS program for the sensitivity analysis the heads are specified at all points on the water table such that

$$H = z \text{ on EFB}$$  \hspace{1cm} (7.5)

and the problem can be solved directly. The approach when using the FREESURF1 program is to specify the heads on the seepage face, EF:

$$H = z \text{ on EF}$$  \hspace{1cm} (7.6)

where $z$ is the water table elevation and the land surface elevation. On the free surface portion of the water table FB the rate of inflow $I$ is defined so that

$$K(x,z) \frac{\partial h}{\partial z} = I \text{ on FB}$$  \hspace{1cm} (7.7)

that is, the water table position is not specified but depends on the amount of precipitation infiltrating the groundwater system. Under these conditions, an iterative solution is necessary.
Hydraulic Conductivity Distribution

The variable conductivities of the idealized geologic environment were discussed in Chapter 6 and depicted in Figure 6.6. Due to the lack of information on the subsurface geology in the Meager area, a number of potential geological configurations were modelled, beginning with the simplest case and working to the most complex.

For the sensitivity analysis only the homogeneous isotropic case was used. The main simulations performed with the FREESURF1 program utilized the four idealized geological configurations illustrated in Figure 7.3. Figure 7.3(a) represents the simplest possible geology. The entire cross-section is assumed to be granodiorite. Figure 7.3(b) includes a fault zone through to exist below Meager Creek. Figure 7.3(c) is a two layer system with the volcanics overlying the granodiorite. Finally, Figure 7.3(d) the most complex geological configuration simulated includes a volcanic layer, granodiorite and a fault zone. The designated impermeable area in Figure 7.3(b and d) is simulated by giving the area a very low hydraulic conductivity (10^{-15} m/s).

Therefore, equipotential lines in Figures 7.7 to 7.11 extend into this zone but insignificant flow occurs there.

The conductivities, anisotropies, and recharge rate values for the system were varied from run to run in an attempt to define the feasible range of parameters for each hydrogeological environment. Feasibility is defined in terms
Figure 7.3 Geologic configurations simulated.
of the flow systems that best fit the calculations presented in this report on the basis of the available field data.

The unconsolidated Quaternary material was not included in the model. The hydraulic conductivity of the valley fill deposits is a number of orders of magnitude larger than the hydraulic conductivity of the basement or volcanics. Therefore, the controlling factor for the amount of water discharging from the system in the hydraulic conductivity distribution in the rock. Furthermore, when water enters the valley fill from the rock it travels down gradient in the same direction as the stream as illustrated in Figure 7.4. The mathematical models used are two dimensional. Consequently, the water movement in the valley fill cannot be modelled.

Finite Element Method

The finite element method works on the premise that a complex function can be approximated piecewise by a finite number of simpler linear functions over small subregions. The subregions are called elements and the corners of the elements are called nodes. The unknown function of hydraulic head within an element can be expressed as a linear function of the heads at the nodes and can be calculated mathematically. It is beyond the scope of this thesis to discuss the finite-element method of solution in detail. The interested reader is referred to Pinder and Gray (1977).

For the FOPS and FREESURF1 programs used here a dif-
Figure 7.4 Movement of groundwater in the basement rock and valley fill.
Different $K_x$ and $K_z$ value can be designated for each element. Head values are specified at each node on a constant head boundary. Recharge rate or total discharge rates are specified for each element on a flux boundary. Impermeable boundaries are handled by the internal operations of the program.

The finite element solution output is the hydraulic head value at each node. A plotting routine in both programs contours the values into a hydraulic head pattern and a flow net can be drawn by hand later, if necessary. The recharge rate and total discharge of water for the elements on the upper boundary are calculated automatically.

Input Data

For both programs the input data required is, 1) a definition of the finite element mesh including nodal coordinates, nodal numbers and element number specifications, 2) one of the geological configurations in Figure 7.3, and 3) the vertical and horizontal hydraulic conductivity values for each of the formations in the configuration.

In the FOPS program the water table position is defined by the programmer and the fluxes across the water table are calculated by the program. The steady-state discharge values, for a given geological configuration are dependent on the water table position chosen. Conversely, in the FREESURF1 program, the flux into the system is defined by the programmer and the position of the water table is calculated
by the program. The steady-state water table position for a given geologic configuration is dependent on the recharge rate chosen.

With an understanding of the simulation strategy at hand, the results of the mathematical modelling can now be discussed.

**Sensitivity Analyses with FOPS**

In Chapter 6, the amount of groundwater discharge out of the Meager Creek basin was calculated. The amount of total discharge depends on the hydraulic conductivity of the material, the water table elevation, and the depth of flow. The program FOPS was used to determine the effect these parameters have on the total discharge. If it is found that a parameter has little effect on the total discharge it can be ignored as a variable input parameter in the FREESURF1 simulations, thereby resulting in a narrowing of the possible range for the other input parameters.

Figure 7.5 illustrates the four different geometries used for the FOPS simulations. FOPS1 represents a flow field 5500 m long extending from -305 m (-1000 ft) to 1525 m (5000 ft) on its right side. The water table decreases in height towards the left to a minimum of 425 m. In FOPS2 the flow field was deepened 610 m while in FOPS3 and FOPS4 the water table was raised above and lowered below the FOPS2 position. Hydraulic conductivity values of $10^{-7}$, $10^{-8}$ and $10^{-9}$ m/s
Figure 7.5 Geometries used in FOPS simulations.
Table 7.1

Summary of Sensitivity Analysis Simulations

<table>
<thead>
<tr>
<th>Geometric Type</th>
<th>Conductivity (m/s)</th>
<th>Total Discharge (m³/s/m)</th>
<th>Maximum Water Table Elevation (m)</th>
<th>Depth of Flow (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>FOP S1</td>
<td>1x10⁻⁷</td>
<td>9.44x10⁻⁶</td>
<td>1525</td>
<td>-305</td>
</tr>
<tr>
<td>FOP S1</td>
<td>1x10⁻⁸</td>
<td>9.44x10⁻⁷</td>
<td>1525</td>
<td>-305</td>
</tr>
<tr>
<td>FOP S1</td>
<td>1x10⁻⁹</td>
<td>9.44x10⁻⁸</td>
<td>1525</td>
<td>-305</td>
</tr>
<tr>
<td>FOP S2</td>
<td>1x10⁻⁷</td>
<td>1.15x10⁻⁵</td>
<td>1525</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S2</td>
<td>1x10⁻⁸</td>
<td>1.15x10⁻⁶</td>
<td>1525</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S2</td>
<td>1x10⁻⁹</td>
<td>1.15x10⁻⁷</td>
<td>1525</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S3</td>
<td>1x10⁻⁷</td>
<td>1.71x10⁻⁵</td>
<td>1830</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S3</td>
<td>1x10⁻⁸</td>
<td>1.71x10⁻⁶</td>
<td>1830</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S3</td>
<td>1x10⁻⁹</td>
<td>1.71x10⁻⁷</td>
<td>1830</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S4</td>
<td>1x10⁻⁷</td>
<td>8.50x10⁻⁶</td>
<td>1220</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S4</td>
<td>1x10⁻⁸</td>
<td>8.50x10⁻⁷</td>
<td>1220</td>
<td>-915</td>
</tr>
<tr>
<td>FOP S4</td>
<td>1x10⁻⁹</td>
<td>8.50x10⁻⁸</td>
<td>1220</td>
<td>-915</td>
</tr>
</tbody>
</table>
where simulated with each geometry. The resulting total discharge is summarized in Table 7.1.

For any given geometry the outflow varies directly with the conductivity of the material. If one increases the conductivity an order of magnitude, then the amount of total discharge increases by an order of magnitude.

For a given hydraulic conductivity value, deepening the flow field 610 m as in geometric type FOPS2 increases the total discharge 1.22 times. This indicates that a very deep flow field would not have a substantially greater total discharge than a similar but shallower system.

When the water table elevation is increased 305 m from 1525 to 1830 m the total discharge increases by 1.5 times. When the water table elevation is lowered 305 m from 1525 to 1220 m the total discharge decreases by 0.5 times. Therefore, the water table must be raised or lowered over large elevation changes in order to cause a substantial change in the total discharge from the system.

It is clear then, that the total discharge from the flow system is very sensitive to hydraulic conductivity values but is reasonably insensitive to water table variations, and especially insensitive to the depth of flow. Accordingly, in the following FREESURF1 simulation results, the hydraulic conductivities and the water table positions will be varied but the depth of flow is chosen arbitrarily and remains cons-
FREESURF1 Modelling

Variable Parameters

The two parameters in Darcy's law that determine the outflow $Q$ are the hydraulic conductivity $K$ and the hydraulic gradient $\frac{dh}{dl}$. The hydraulic gradients depend on the hydraulic head field, and in the system modelled in this report the head field depends on the position of the water table which in turn depends on the amount of inflow $I$ or recharge rate into the geologic system. The combination of the variables $I$ and $K$ will determine the flux $Q$ out of the system.

In Chapter 6 it was determined that the contribution of groundwater flow in the Meager basin above the stage location is in the range 2.5 m$^3$/s. The Meager Mountain side of the river constitutes 45% of the surface area of the basin; therefore we will assume that 45% of the total groundwater flow (1.13 m$^3$/s) is contributed from this side. It enters over the 18,500 m length of the valley. The two-dimensional cross-section model has a thickness of 1 m, consequently its contribution to the total total discharge is $6\times10^{-3}$ m$^3$/s/m. We will designate this flux per unit length as $q$; it has units of m$^2$/s.

The recharge rate was estimated previously at 0.36 m/y or $1.14x10^{-8}$ m/s. The above estimated values of $I$ and $q$
allow for the calculation of different feasible hydraulic conductivity distributions using FREESURF1.

Finite Element Mesh

The finite element mesh used for the FREESURF1 simulations is illustrated in Figure 7.6. It is superimposed on a simplified version of the geology by Fairbank et al. (1979). The quadrilateral elements above 600 m elevation are collapsible. They can flex upwards and downwards to allow the water table to rise or lower to the steady-state position. The intention of the author was to simulate the volcanics as in Figure 7.6, however the program has the limitation that all the elements in the collapsible part of the mesh have to possess the same hydraulic conductivity. For this reason the volcanic layers in Figure 7.3(c) and (d) are more extensive than field mapping and drill holes indicate. It is felt that this will cause only minor inaccuracies in the simulations.

The program also has limitations situating the water table in its exact location when there is an irregular ground surface as in these simulations. The simulated position of the water table may be slightly in error at the discharge end of the system but it is felt that the possible errors will have negligible effect on the flow field.
Figure 7.6 Finite element mesh used in FREESURF1 simulations.

Gd - Granodiorite
P1 - Rhyodacite tuff, breccia and flows
P3 - Porphyritic andesite flows
P8 - Porphyritic rhyodacite flows
Results

For the simulations the hydraulic conductivity distributions were selected as in Table 7.2. The recharge rate into the system was varied until the maximum and minimum water table configuration were found. The resulting contoured hydraulic head fields are depicted in Figures 7.7 to 7.11. The maximum value occurs when the water table intersects the ground surface in the highland valleys between mountain peaks as in Figure 7.7(a). The minimum occurs when the elevation of node 252 becomes less than the elevation of node 271 (Figure 7.6). Node 252 is the last node that is moveable. Node 271 is set at a constant head equal to the elevation of Meager Creek. When node 252 moves lower than 271 it indicates the water table elevation is too low to be physically realistic.

The lines in Figures 7.7 to 7.11 which join points of equal hydraulic head are called equipotential lines. For the isotropic cases flowlines can be drawn perpendicular to the equipotential lines to construct a flow net as in Figure 7.7(b). For the anisotropic cases flowlines are not necessarily perpendicular to the equipotential lines. There are graphic methods for drawing the flowlines but they cannot be drawn on the diagrams directly. The R/D designations on the diagrams indicate the boundary between the recharge and discharge areas. The dashed lines show the geological boundaries as per Figure 7.3. The hydraulic conductivities used
### Table 7.2

Hydraulic Conductivity Distributions

<table>
<thead>
<tr>
<th>Geological Configuration (as in Fig. 7.3)</th>
<th>Simulation Designation</th>
<th>Conductivity Distribution (m/s)</th>
<th>Dimensionless Ratios</th>
</tr>
</thead>
<tbody>
<tr>
<td>FREESURF 1</td>
<td>A</td>
<td>$K_g = 10^{-8}$</td>
<td></td>
</tr>
<tr>
<td>FREESURF 1</td>
<td>B</td>
<td>$K_{gx} = 10^{-7}$ $K_{gz} = 10^{-8}$</td>
<td>$K_{gx} = 10K_{gz}$</td>
</tr>
<tr>
<td>FREESURF 1</td>
<td>C1</td>
<td>$K_{gx} = 10^{-8}$ $K_{gz} = 10^{-7}$</td>
<td>$K_{gz} = 10K_{gx}$</td>
</tr>
<tr>
<td>FREESURF 1</td>
<td>C2</td>
<td>$K_{gx} = 5 \times 10^{-9}$ $K_{gz} = 5 \times 10^{-8}$</td>
<td>$K_{gz} = 10K_{gx}$</td>
</tr>
<tr>
<td>FREESURF 2</td>
<td>A</td>
<td>$K_{g} = 10^{-7}$ $K_{f} = 10^{-15}$ $K_{b} = 10^{-15}$</td>
<td>$K_{g} = 10K_{g}$ &amp; $K_{f} = 2K_{g}$</td>
</tr>
<tr>
<td>FREESURF 2</td>
<td>B</td>
<td>$K_{gx} = 5 \times 10^{-9}$ $K_{gz} = 5 \times 10^{-8}$ $K_{f} = 10^{-7}$ $K_{b} = 10^{-15}$</td>
<td>$K_{gz} = 10K_{gx}$</td>
</tr>
<tr>
<td>FREESURF 3</td>
<td>A</td>
<td>$K_v = 10^{-7}$ $K_g = 10^{-8}$</td>
<td>$K_v = 10K_g$</td>
</tr>
<tr>
<td>FREESURF 3</td>
<td>B</td>
<td>$K_v = 5 \times 10^{-8}$ $K_{gx} = 5 \times 10^{-9}$ $K_{gz} = 5 \times 10^{-8}$</td>
<td>$K_{gz} = K_v = 10K_{gx}$</td>
</tr>
<tr>
<td>FREESURF 4</td>
<td>A</td>
<td>$K_v = 5 \times 10^{-8}$ $K_f = 10^{-7}$ $K_g = 10^{-8}$ $K_b = 10^{-15}$</td>
<td>$K_f = 10K_g = 2K_v$</td>
</tr>
<tr>
<td>FREESURF 4</td>
<td>B</td>
<td>$K_v = 5 \times 10^{-8}$ $K_f = 10^{-7}$ $K_{gx} = 5 \times 10^{-9}$ $K_{gz} = 5 \times 10^{-8}$</td>
<td>$K_{gz} = K_v = 10K_{gx}$ &amp; $K_f = 2K_g$</td>
</tr>
</tbody>
</table>

$K_g$ = conductivity of the granodiorite.

$K_{gx}$ = conductivity of the granodiorite in the horizontal direction.

$K_{gz}$ = conductivity of the granodiorite in the vertical direction.

$K_v$ = conductivity of the volcanics.

$K_f$ = conductivity of the fault.

$K_b$ = conductivity of the impermeable area.
Figure 7.7 Equipotential pattern for FREESURF 1A and FREESURF 1B cases.
Figure 7.8 Equipotential pattern for FREESURF 1C cases.
Figure 7.9 Equipotential pattern for FREESURF 2 cases.
Figure 7.10 Equipotential pattern for FREESURF 3 cases.
Figure 7.11 Equipotential pattern for FREESURF 4 cases.
for each geologic unit can be gleaned from Table 7.2 while relative horizontal and vertical hydraulic conductivities are illustrated on the equipotential diagrams.

Case 1A is homogeneous and isotropic; cases 1B and 1C are homogeneous and anisotropic. A comparison of these three cases illustrates the changes in the equipotential field caused by anisotropy. Note in the diagrams of FREESURF 1B that the flow field becomes distorted due to a greater horizontal anisotropy. A larger horizontal hydraulic conductivity than vertical is unrealistic for the basement rock but is shown here to display the changes in the equipotential field caused by anisotropy.

In the 1C1 vs 1C2 cases the hydraulic conductivity is changed but the ratio between the horizontal and vertical hydraulic conductivity, $K_X = K_Z$ remains constant at 1:10. Note in Figure 7.8 that the hydraulic head field in the two cases is nearly identical. Therefore, as long as the ratios between the various hydraulic conductivity values remains constant and the recharge rate is increased or decreased the same amount as the conductivities, the flow pattern will not change.

FREESURF 2 displays the effect of a more permeable fault in the system as shown in Figure 7.3 (b). The partial flow net in Figure 7.9(b) shows how the fault acts as a highway to the surface for fluids that intersect the zone. If the hypothetical fault does exist it would become tighter and much
less conductive at depth, analogous to faults discussed in Chapter 5 studied by Raven and Gale (1977). Most fault systems have clay seams or slickensides that act as an impermeable boundary for fluids so the area below the fault is given a very low conductivity to simulate this impermeable boundary.

Figure 7.10(a and b) are the FREESURF 3A equipotential patterns for isotropic volcanics over isotropic basement rock. Recall from Darcy's law that the discharge through the rock depends on the hydraulic conductivity $K$ and the hydraulic gradient $\frac{dh}{dl}$. The hydraulic gradient in the volcanics and basement rock is similar but the hydraulic conductivity is one order of magnitude larger in the volcanics than in the basement rock. Therefore, most of the flux will be through the volcanics. As stated earlier this situation would lead to year round spring lines at the basal volcanic contact and few exist at Meager Mountain. Figure 7.10(c and d) represents a more likely situation with the volcanics and basement rock having the same vertical conductivity. The equivalent isotropic hydraulic conductivity of the basement rock can be calculated using the formula $K' = \sqrt{K_xK_z}$, where $K'$ is the equivalent isotropic hydraulic conductivity, $K_x$ is the horizontal hydraulic conductivity, and $K_z$ is the vertical hydraulic conductivity of the basement rock. This value calculates to be $1.58 \times 10^{-8}$ or 3 times less than that of the volcanics. This small difference would not substantially restrict the flow from the volcanics into the basement.
Figures 7.11 (a and b) represent the FREESURF 4A equipotential patterns for isotropic volcanics over isotropic basement with a fault in the basement rock extending towards the Meager valley. The hydraulic conductivity of the volcanics is five times greater than the basement and the hydraulic conductivity of the fault is 10 times that of the basement. Again, this configuration would cause spring lines at the volcanic-basement contact, and few occur.

The FREESURF 4B simulations are the most realistic, if a fault exists. The equipotential patterns in Figure 7.11(c and d) represent the situation where isotropic volcanics overlie anisotropic basement rocks with an isotropic permeable fault zone. The vertical hydraulic conductivity of the basement granodiorite and the volcanics is the same. The horizontal conductivity of the volcanics is greater than that of the granodiorite and the most permeable zone is the fault area.

The fluid pressure $P$ is related to the pressure head as shown in Equation 3.6. Therefore, the pressure at any point in the equipotential diagrams in Figures 7.7 to 7.11 can be calculated. Figure 7.12 and 7.13 are pressure vs depth graphs for the various simulations. The values where taken vertically downward below the unconsolidated-bedrock contact as shown by the line A-B in Figure 7.7(d). This position was chosen because it is the area where a number of holes have been drilled. In the minimum water table elevation examples,
Figure 7.12 Pressure vs depth graph of minimum water table elevation examples.
Figure 7.13 Pressure vs depth graph of maximum water table elevation examples.
the bottom hole pressure varies within plus or minus 500 Pa of hydrostatic pressure. In the maximum water table elevation examples, the bottom hole pressure varies within plus or minus 900 Pa of isostatic pressure. In other words, at 900 m the pressure only varies within 5 to 10% of isostatic pressure in any of the geologic configurations modelled.

The graphs in Figures 7.7 to 7.11 can be considered dimensionless if the ratios of the different hydraulic conductivity distributions remain constant. The ratios for each case simulated are tabulated in Table 7.2. In other words, if the hydraulic conductivities are doubled the recharge rate must be doubled to attain the same water table elevations and equipotential pattern. The consequence of this, of course, is that the total discharge also doubles.

Of the three variable parameters, hydraulic conductivity, recharge rate, and total discharge, the total discharge is the most accurately estimated at approximately $6 \times 10^{-5}$ m$^3$/s. The total discharge of the original simulations can be multiplied by a factor to produce a total discharge of $6 \times 10^{-5}$ m$^3$/s. To keep the same water table configuration for each graph the conductivities and recharge rate must also be multiplied by the same factor. The end result is summarized in Table 7.3. The table illustrates that with the set total discharge, the recharge rate needed varies between $1.09 \times 10^{-4}$ and $1.81 \times 10^{-4}$ or 14 and 23% of the total precipitation. The water balance calculations of Chapter 4 and estimates of
Table 7.3

Influx and Hydraulic Conductivity Distributions with a Set Outflux

<table>
<thead>
<tr>
<th>Type</th>
<th>Recharge Rate (m/s)</th>
<th>Total Discharge (m³/s/m)</th>
<th>Conductivity (m/s)</th>
<th>% Total Precipitation</th>
</tr>
</thead>
<tbody>
<tr>
<td>IA Max Flux</td>
<td>1.77x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=9.12x10⁻⁸</td>
<td>22.3</td>
</tr>
<tr>
<td>IA Min Flux</td>
<td>1.09x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=1.52x10⁻⁷</td>
<td>13.8</td>
</tr>
<tr>
<td>IB Max Flux</td>
<td>1.76x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=2.78x10⁻⁷, Kᵣz=2.78x10⁻⁸</td>
<td>22.2</td>
</tr>
<tr>
<td>IB Min Flux</td>
<td>1.09x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=4.23x10⁻⁷, Kᵣz=4.23x10⁻⁸</td>
<td>13.8</td>
</tr>
<tr>
<td>IC Max Flux</td>
<td>1.76x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=5.94x10⁻⁷, Kᵣz=5.94x10⁻⁷</td>
<td>22.2</td>
</tr>
<tr>
<td>IC Min Flux</td>
<td>1.01x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=6.91x10⁻⁷, Kᵣz=6.91x10⁻⁷</td>
<td>12.7</td>
</tr>
<tr>
<td>IC Max Flux</td>
<td>1.81x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=6.15x10⁻⁷, Kᵣz=6.15x10⁻⁷</td>
<td>22.8</td>
</tr>
<tr>
<td>IC Min Flux</td>
<td>1.19x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=7.32x10⁻⁷, Kᵣz=7.32x10⁻⁷</td>
<td>13.9</td>
</tr>
<tr>
<td>2 Max(iso)</td>
<td>1.78x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=7.41x10⁻⁷, Kᵣz=7.41x10⁻⁷</td>
<td>22.5</td>
</tr>
<tr>
<td>2 Min(iso)</td>
<td>1.09x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=1.04x10⁻⁷, Kᵣz=1.04x10⁻⁷</td>
<td>13.8</td>
</tr>
<tr>
<td>2 Max(aniso)</td>
<td>1.79x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=9.84x10⁻⁷, Kᵣz=4.92x10⁻⁷, Kᵣ₃=4.92x10⁻⁷</td>
<td>22.6</td>
</tr>
<tr>
<td>2 Min(aniso)</td>
<td>1.11x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=9.02x10⁻⁷, Kᵣz=4.51x10⁻⁷, Kᵣ₃=4.92x10⁻⁷</td>
<td>14.0</td>
</tr>
<tr>
<td>3 Max(iso,gran)</td>
<td>1.77x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=1.62x10⁻⁷, Kᵣz=1.62x10⁻⁷</td>
<td>22.3</td>
</tr>
<tr>
<td>3 Min(iso,gran)</td>
<td>1.09x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=2.99x10⁻⁷, Kᵣz=2.99x10⁻⁷</td>
<td>13.8</td>
</tr>
<tr>
<td>3 Max(aniso,gran)</td>
<td>1.77x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=1.63x10⁻⁷, Kᵣz=1.63x10⁻⁷, Kᵣ₃=1.63x10⁻⁷</td>
<td>22.3</td>
</tr>
<tr>
<td>3 Min(aniso,gran)</td>
<td>1.09x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=2.71x10⁻⁷, Kᵣz=2.71x10⁻⁷, Kᵣ₃=2.71x10⁻⁷</td>
<td>13.8</td>
</tr>
<tr>
<td>4 Max(iso)</td>
<td>1.76x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=1.36x10⁻⁷, Kᵣz=2.71x10⁻⁷, Kᵣ₃=2.71x10⁻⁷</td>
<td>22.2</td>
</tr>
<tr>
<td>4 Min(iso)</td>
<td>1.09x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=2.22x10⁻⁷, Kᵣz=4.44x10⁻⁷, Kᵣ₃=4.44x10⁻⁷</td>
<td>13.8</td>
</tr>
<tr>
<td>4 Max(aniso)</td>
<td>1.77x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=1.42x10⁻⁷, Kᵣz=2.83x10⁻⁷, Kᵣ₃=1.42x10⁻⁷</td>
<td>22.3</td>
</tr>
<tr>
<td>4 Min(aniso)</td>
<td>1.10x10⁻⁸</td>
<td>6x10⁻⁵</td>
<td>Kᵣ=2.24x10⁻⁷, Kᵣz=4.48x10⁻⁷, Kᵣ₃=2.24x10⁻⁷</td>
<td>13.9</td>
</tr>
</tbody>
</table>

Kᵣ = Conductivity of the granodiorite
Kᵣₓ = Conductivity of the granodiorite in the horizontal direction
Kᵣ₃ = Conductivity of the granodiorite in the vertical direction
Kᵥ = Conductivity of the volcanics
Kᵣ = Conductivity of the fault
Kᵣᵦ = Conductivity of the impermeable boundary
groundwater recharge in Chapter 6 indicated that 13 to 16% of the total precipitation became groundwater flow. The model values are thus in the same range but slightly higher.

Note in Figures 7.7 to 7.11 the position of the discharge and recharge areas. The maximum water table elevation examples indicate that the discharge area is situated a long way up the mountain side although no sign of the this was observed in the field. As mentioned in Chapter 6 the discharge area is probably restricted to the valley bottom covered by unconsolidated valley fill deposits as in the minimum examples. Intermediate simulations between the maximum and minimum extremes show that water table elevations as low as the minimum position or as high as about halfway between the minimum and maximum positions have the same acceptable discharge area. This means that the range of feasible percentages of the total precipitation entering the groundwater system drops from 14 to 22% down to 14-18% which corresponds very well with the 13-16% range estimated from the water balance and the estimates of groundwater recharge.

Interpretation

In the following discussion the unrealistic cases 1B, 3A and 4A will not be included. Table 7.3 illustrates that in any geologic unit the highest value of conductivity in the table is only one fifth to one order of magnitude larger than the lowest value. When one considers that conductivities can
vary over 10 orders of magnitude or more, it is clear that the modelling has produced a very definitive set of conductivities for the different units. The horizontal hydraulic conductivity of the basement granodiorite $K_x$ varies from $1.42 \times 10^{-8}$ to $1.52 \times 10^{-7}$ m/s while in the vertical direction it varies from $7.41 \times 10^{-8}$ to $7.32 \times 10^{-7}$ m/s. The hydraulic conductivity of the volcanics $K_v$ ranges from $1.42 \times 10^{-7}$ to $2.71 \times 10^{-7}$ m/s and in the permeable fault it ranges from $2.82 \times 10^{-7}$ to $1.04 \times 10^{-6}$ m/s.

It must be stressed, however, that these answers are only for the conductivity ratios in Table 7.2. Theoretically there are an infinite number of possible conductivity distributions and therefore, an infinite conductivity range for each geologic formation. Nevertheless, the geologic environments, recharge rate values and total discharge values chosen or calculated are reasonable and the author feels that the hydraulic conductivities in Table 7.3 are within an order of magnitude of the true values. Even though the geology is not as simple as depicted in the models, the simulations show that the conductivity of the granodiorite varies only slightly from the simplest to the most complex geology leading one to believe that the values are realistic.

In conclusion, from the geological configurations chosen, the modelling indicates that the granodiorite has a horizontal hydraulic conductivity of approximately $5 \times 10^{-8}$ m/s and a vertical hydraulic conductivity of $1 \times 10^{-7}$ m/s. The
hydraulic conductivity of the volcanics is close to $1 \times 10^{-7}$ m/s, similar to the vertical component in the granodiorite and the most permeable zone, the fault, has a hydraulic conductivity of roughly $7 \times 10^{-7}$ m/s.
Chapter 8.

CONCLUSIONS AND RECOMMENDATIONS

In this thesis an attempt has been made to describe the hydrogeology of the Meager Mountain geothermal area. The main purpose of the thesis was to develop a preliminary, mathematical model of the regional groundwater flow in the area. A summary of the major findings of this research is presented below.

Summary and Conclusion

Geography

1. The topography in the Meager Mountain area is very rugged. The relief ranges from 425 to 2700 m over horizontal distances of roughly 5 km. Extensive glacier systems terminate at the headwaters of a number of high-gradient, youthful streams. The streams produce a radial pattern of drainage around the mountain.

2. Precipitation in the coast mountains is variable but a total annual precipitation rate of 2.5 m/y is expected at Meager Mountain.

3. The mean annual runoff of the Lillooet River is 1880 mm/yr while runoff measurements at Miller Creek which is at a higher elevation are estimated to be 2400 ± 300 mm/yr.

4. The potential evapotranspiration in the Meager Mountain area is approximately 500 mm/y.
5. The Meager Creek hotsprings and Pebble Creek hotsprings have a combined discharge rate of approximately 45 l/s. They represent a minute fraction of the regional groundwater discharge.

6. A number of cold springs exist throughout the Meager Mountain area. Most have discharges less than 5 l/s. They represent about 1 to 2% of the regional groundwater flow.

Geology

1. Meager Mountain is part of the Garibaldi belt of north-south trending Quaternary volcanoes. Meager Mountain initially erupted in the Pliocene and has been dormant for approximately the past 2500 years. The older portion of the mountain comprises mainly widespread andesite and is best exposed in the south. The younger north half is composed of dacite flows and lava domes overlying the older andesite flows.

2. The volcanics erupted through a basement consisting of Tertiary and older granitic rocks. The explosive nature of the volcanic eruptions severely may have fractured the basement rock thereby increasing its fracture permeability.

3. The Lillooet River and Meager Creek valleys are filled with unconsolidated deposits varying from zero to over 250 m in thickness. The deposits consist of sand and gravels with interbedded till and lacustrine clay layers.
Hydrogeology

1. The preliminary water balance undertaken for the entire Lillooet basin indicates that 17% of the precipitation enters the groundwater system.

2. The Meager Mountain fracture survey indicates two dominant, near vertical joint sets, however, the study did not supply sufficient data for the calculation of the rock mass hydraulic conductivity. It is the author's opinion that there is no suitable method to make surface measurements of fractured rocks in the field to attain a representative estimate of hydraulic conductivity for the rock mass at depth.

3. The study of published fracture permeabilities of various rock types revealed that the hydraulic conductivity of most fractured rock lies in the $10^{-7}$ to $10^{-11}$ m/s range. It is felt that the hydraulic conductivity of the basement granodiorites in the study area is at the high end of this range, due to their highly fractured nature. Values of $10^{-7}$ to $10^{-9}$ m/s could be expected.

4. It can be shown that the most likely position for the water table is at an intermediate elevation in the mountain system. Apart from a very few springs at higher elevations, the discharge area is believed to be confined to the section of the valley overlain by unconsolidated deposits.

5. The fracture hydraulic conductivity that occurs across all the volcanic layers has greater control over the
groundwater flow than the intergranular hydraulic conductivity differences between layers. In this study, the entire volcanic pile is considered as one hydrogeologic unit. The basement rock and the unconsolidated valley fill are two other distinct hydrogeologic units in the study area.

6. The representative hydraulic conductivities as determined by published data and field observations are estimated to be $10^{-5}$ to $10^{-3}$ m/s for the unconsolidated deposits, $10^{-7}$ to $10^{-4}$ m/s for the basement rock and $10^{-11}$ to $10^{-8}$ m/s for the volcanics. The hydraulic conductivity of the basement rock is perhaps as much as 5 times greater in the vertical direction than in the horizontal direction due to the extensive vertical fracturing. The hydraulic conductivity of the volcanic rock is perhaps as much as 5 times greater in the horizontal direction than in the vertical direction due to layering. The vertical hydraulic conductivity in the basement rock is similar to that in the volcanic rock.

7. The groundwater discharge in the Meager Creek basin calculates to be 14.5% of the total precipitation, assuming an annual rainfall rate of 2.5 m/y. This discharge value range correlates well with the 17% calculated for the entire Lillooet basin through the use of a water balance.
Mathematical Modelling

1. The parameters that must be known to mathematically model the groundwater flow in an area include, 1) the region of flow, 2) the equation of flow within the region, 3) the boundary conditions around the boundaries of the region, 4) the spatial distribution of hydraulic conductivity values in the region.

2. The sensitivity analysis using the program FOPS reveals that the amount of discharge varies directly with the hydraulic conductivity, undergoes moderate changes with water table elevation variations and is insensitive to depth of flow variations. Consequently, in the main simulations the hydraulic conductivity and water table elevations are varied but the depth of flow is chosen arbitrarily and remains constant.

3. The geochemical studies of the thermal waters at Meager Mountain completed by Hammerstrom and Brown (1977) and Clark (1980) indicate that there is no evidence that the water temperature ever exceeded 140°C. This indicates that the flow field for the thermal waters probably is a shallow one. The mathematical modelling cannot prove that it is a shallow flow system, however it illustrates that discharge out of the system is independent of the depth of flow. Therefore, a shallow flow system is feasible and the geochemical studies indicate it is the most likely situation.

4. The program FREESURF1 was used for the main simulations
in this report. The most accurately known variable parameter in the Meager Mountain area is the discharge. In the simulations, the discharge value is held constant and a possible range of water table elevations and hydraulic conductivity distributions are calculated for the 4 geological configurations considered. The simulations determined that 14-18% of the total precipitation enters the groundwater zone which correlates well with the value of 17% determined in the water balance studies, and the value of 14.5% determined in the Meager Creek baseflow studies.

5. The bottom hole fluid pressure at 900 m in any of the simulations varies within 5 to 10% of isostatic pressure.

6. The mathematical simulations indicated the most likely values for the hydraulic conductivity to be $5 \times 10^{-8}$ m/s in the horizontal direction in the basement rocks, $1 \times 10^{-7}$ for the vertical direction in the basement rock, $1 \times 10^{-7}$ for the volcanics and $7 \times 10^{-7}$ for the fault zone. Within one order of magnitude of the actual values.

**Recommendations**

An accurate estimate of the amount of groundwater discharge in the Meager Creek basin will allow a more precise overall hydraulic conductivity to be assessed for the basement rock. A detailed water balance is recommended to attain this discharge value. The water balance would have to be performed over a number of years and would involve the fol-
lowing measurements: 1) Measure the volume change of the glaciers in the basin area to determine their contribution to basin discharge. 2) Measure the precipitation from the valley floor to the higher elevations to ascertain the annual precipitation and the change of precipitation values with elevation. 3) Measure the discharge year-round at the stage sites currently in use and install another on the Lillooet River just south of its confluence with Meager Creek to calculate the total discharge out of the area and sub-basin discharges. 4) Measure the evaporation by using pan data and solar radiation data to estimate the evapotranspiration in the area.

A fracture survey in the volcanics would give the general trends of fracturing. The survey might not be useful for hydraulic conductivity calculations but would determine whether the fractures are mainly vertical as stated in the hydrogeologic interpretation of this report and possibly reveal zones of relatively higher fracture density to possibly reveal zones of higher hydraulic conductivity at depth.

A more refined geological configuration and hydraulic conductivity distribution for the basement rock will evolve with continued drilling and testing. As these two variable parameters become more refined, mathematical simulations of the hydrogeologic environment will also yield more refined results. Mathematical modelling will definitely be of value to this project throughout its exploration, development and production stages.
The two limiting factors for development of the Meager Mountain geothermal project are the hydraulic conductivity of the basement rock and the temperature of the rock at depth. Temperature logging of the drill holes has revealed very promising geothermal gradients. Hydraulic conductivity tests have not been undertaken. It is imperative that down hole conductivity tests be performed from the top to the bottom of all holes to reveal the hydraulic conductivity distribution spatially and with depth. Near surface, the hydraulic conductivity may be sufficiently high for production but the temperature may be too low. At depth, the temperature may be sufficiently high for production but the hydraulic conductivity may be too low.

The author feels that a change of strategy is in order for the Meager Mountain geothermal project. In the initial stages, exploration geophysics and drilling are necessary to outline the promising region for development. Now that this region has been delineated in the South Reservoir area a more sophisticated technology needs to be employed. Numerous companies exist that perform specialized down hole hydrogeologic tests on fractured rock. The employment of these companies is very expensive, however the information they can generate is essential to the project at this time.
APPENDIX I

GLACIAL BASAL MELT FLUX CALCULATIONS
IN THE MEAGER CREEK BASIN
The governing equation for the rate of glacial basal melt due to the earth's geothermal flux is

\[ \frac{dz}{dt} = \frac{q_0}{\rho L} \]

where \( \frac{dz}{dt} \) is the rate of change of thickness of ice with time, \( q_0 \) the geothermal heat flux (W/m\(^2\)), \( \rho \) the density of ice (Kg/m\(^3\)) and \( L \) the latent heat of melting (J/Kg). The world average heat flux is 0.05 W/m\(^2\) and the density of ice and latent heat of melt have values of 900 Kg/m\(^3\) and 3.35x10\(^{-5}\) J/Kg respectively.

In the Meager Mountain area varies from 0.1 to 0.93 W/m\(^2\) (Lewis and Souther, 1978) with an average of 0.5 W/m\(^2\). This is ten times the world average. The rate of change of thickness of the glaciers with time in the Meager Mountain area is therefore

\[ \frac{dz}{dt} = 1.66 \times 10^{-9} \text{ m/s} \]

The area of the glaciers contributing to the flow in Meager Creek is approximately 51.4 Km\(^2\) or 5.14x10\(^7\) m\(^2\). This figure was calculated using a planimeter and topography map. The volume \( Q \) of fluid release can now be calculated as

\[ Q = \frac{dv}{dt} = A \frac{dz}{dt} = 0.09 \text{ m}^3/\text{s} \]

The measured groundwater flow out of the area is 2.5 m\(^3\)/s; therefore the contribution of basal glacial melt to runoff in the winter months should not exceed 3 to 4%.
APPENDIX II

CONTOURED-POLE PLOTS OF FRACTURE DATA

- Joint set #1
- Joint set #2
- Joint set #3
- Joint set #4
GEODAT - LOWE HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SR01 : 5 602 650 N 461 000 E 1100 ft.
OBSERVATIONS: 190
POPULATION: 190

GEODAT - LOWE HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SR02 : 5 601 650 N 466 200 E 760 ft.
OBSERVATIONS: 133
POPULATION: 133
CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SR03 : 5 602 300 N 466 600 E 760 #1.
OBSERVATIONS: 200
POPULATION: 200

CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SR04 : 5 601 900 N 462 700 E 850 #1.
OBSERVATIONS: 99
POPULATION: 99
CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SR07: 5 662 000 N: 462 500 E: 1000 s1.
OBSERVATIONS: 199
POPULATION: 199

CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SR08: 5 662 200 N: 467 950 E: 740 s1.
OBSERVATIONS: 250
POPULATION: 250
GEOHAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SR9B-9A : 5 601 000 N : 463 075 E : 700 #1.
OBSERVATIONS: 348
POPULATION: 348

GEOHAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
OBSERVATIONS: 216
POPULATION: 216
GEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
OBSERVATIONS: 170
POPULATION: 170

GEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
OBSERVATIONS: 149
POPULATION: 149
CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT

CONTOUR PLOT FOR: SITE SRI5 : 5 600 650 N : 461 650 E : 760 #1.
OBSERVATIONS: 103
POPULATION: 103

CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT

OBSERVATIONS: 101
POPULATION: 101
GEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT

OBSERVATIONS: 107
POPULATION: 107

GEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT

OBSERVATIONS: 182
POPULATION: 182
CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT

CONTOUR PLOT FOR: SITE SR21: 5 601 500 N : 462 630 E : 930±1
OBSERVATIONS: 151
POPULATION: 151

CEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT

CONTOUR PLOT FOR: SITE SR22: 5 601 225 N : 467 750 E : 900±1
OBSERVATIONS: 200
POPULATION: 200
GEODAT - LOWER HEMISPHERE EQUAL AREA POLAR PLOT
CONTOUR PLOT FOR: SITE SF-25 : S. 60° 675 H : 466 925 E : 1298 #1.
OBSERVATIONS: 148
POPULATION: 148
APPENDIX III

RIVER STAGE MEASUREMENTS AT
MEAGER CREEK HOTSPRINGS BRIDGE, 1979.
<table>
<thead>
<tr>
<th>Date</th>
<th>Time</th>
<th>Water Level (in Feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Observed</td>
</tr>
<tr>
<td>June</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>1545</td>
<td>152.58</td>
</tr>
<tr>
<td>19</td>
<td>0930</td>
<td>152.58</td>
</tr>
<tr>
<td>25</td>
<td>0910</td>
<td>152.43</td>
</tr>
<tr>
<td></td>
<td>1235</td>
<td>152.58</td>
</tr>
<tr>
<td>July</td>
<td></td>
<td></td>
</tr>
<tr>
<td>03</td>
<td>1525</td>
<td>152.08</td>
</tr>
<tr>
<td>04</td>
<td>1510</td>
<td>152.25</td>
</tr>
<tr>
<td>05</td>
<td>1650</td>
<td>152.65</td>
</tr>
<tr>
<td>06</td>
<td>1700</td>
<td>152.78</td>
</tr>
<tr>
<td>07</td>
<td>1515</td>
<td>152.63</td>
</tr>
<tr>
<td>08</td>
<td>1510</td>
<td>153.8</td>
</tr>
<tr>
<td>09</td>
<td>1910</td>
<td>153.34</td>
</tr>
<tr>
<td>10</td>
<td>1930</td>
<td>153.81</td>
</tr>
<tr>
<td>11</td>
<td>1820</td>
<td>153.23</td>
</tr>
<tr>
<td>12</td>
<td>1930</td>
<td>153.13</td>
</tr>
<tr>
<td>13</td>
<td>1900</td>
<td>152.7</td>
</tr>
<tr>
<td>14</td>
<td>2120</td>
<td>153.08</td>
</tr>
<tr>
<td>15</td>
<td>2100</td>
<td>153.18</td>
</tr>
<tr>
<td>16</td>
<td>1130</td>
<td>152.73</td>
</tr>
<tr>
<td>17</td>
<td>2025</td>
<td>153.68</td>
</tr>
<tr>
<td>18</td>
<td>2045</td>
<td>153.88</td>
</tr>
<tr>
<td>19</td>
<td>2030</td>
<td>153.88</td>
</tr>
<tr>
<td>20</td>
<td>0900</td>
<td>152.88</td>
</tr>
<tr>
<td>21</td>
<td>1825</td>
<td>153.68</td>
</tr>
<tr>
<td>22</td>
<td>1830</td>
<td>153.43</td>
</tr>
<tr>
<td>23</td>
<td>1915</td>
<td>153.48</td>
</tr>
<tr>
<td>24</td>
<td>2030</td>
<td>153.28</td>
</tr>
<tr>
<td>25</td>
<td>1600</td>
<td>153.28</td>
</tr>
<tr>
<td>26</td>
<td>1900</td>
<td>153.38</td>
</tr>
<tr>
<td>27</td>
<td>1605</td>
<td>153.33</td>
</tr>
<tr>
<td>28</td>
<td>1100</td>
<td>152.83</td>
</tr>
<tr>
<td>29</td>
<td>1630</td>
<td>153.03</td>
</tr>
<tr>
<td>30</td>
<td>1535</td>
<td>153.13</td>
</tr>
<tr>
<td>31</td>
<td>1915</td>
<td>153.33</td>
</tr>
<tr>
<td>Date 1979</td>
<td>Time</td>
<td>Observed</td>
</tr>
<tr>
<td>----------</td>
<td>-------</td>
<td>----------</td>
</tr>
<tr>
<td>August</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>1635</td>
<td>153.28</td>
</tr>
<tr>
<td>2</td>
<td>1530</td>
<td>153.13</td>
</tr>
<tr>
<td>3</td>
<td>1645</td>
<td>153.08</td>
</tr>
<tr>
<td>4</td>
<td>1645</td>
<td>152.98</td>
</tr>
<tr>
<td>5</td>
<td>1615</td>
<td>152.83</td>
</tr>
<tr>
<td>6</td>
<td>1420</td>
<td>152.48</td>
</tr>
<tr>
<td>7</td>
<td>1900</td>
<td>152.86</td>
</tr>
<tr>
<td>8</td>
<td>1630</td>
<td>153.18</td>
</tr>
<tr>
<td>9</td>
<td>1930</td>
<td>153.39</td>
</tr>
<tr>
<td>10</td>
<td>1530</td>
<td>153.14</td>
</tr>
<tr>
<td>11</td>
<td>1920</td>
<td>153.45</td>
</tr>
<tr>
<td>12</td>
<td>1810</td>
<td>153.59</td>
</tr>
<tr>
<td>13</td>
<td>1915</td>
<td>153.68</td>
</tr>
<tr>
<td>14</td>
<td>1825</td>
<td>154.25</td>
</tr>
<tr>
<td>15</td>
<td>2030</td>
<td>153.25</td>
</tr>
<tr>
<td>16</td>
<td>1940</td>
<td>153.31</td>
</tr>
<tr>
<td>17</td>
<td>1700</td>
<td>153.25</td>
</tr>
<tr>
<td>18</td>
<td>1900</td>
<td>153.19</td>
</tr>
<tr>
<td>19</td>
<td>1845</td>
<td>153.23</td>
</tr>
<tr>
<td>20</td>
<td>2100</td>
<td>153.55</td>
</tr>
<tr>
<td>21</td>
<td>1845</td>
<td>153.78</td>
</tr>
<tr>
<td>22</td>
<td>1920</td>
<td>153.73</td>
</tr>
<tr>
<td>23</td>
<td>2035</td>
<td>153.57</td>
</tr>
<tr>
<td>24</td>
<td>1900</td>
<td>153.31</td>
</tr>
<tr>
<td>25</td>
<td>1915</td>
<td>153.48</td>
</tr>
<tr>
<td>26</td>
<td>1910</td>
<td>153.58</td>
</tr>
<tr>
<td>27</td>
<td>1130</td>
<td>152.95</td>
</tr>
<tr>
<td></td>
<td>2015</td>
<td>153.34</td>
</tr>
<tr>
<td>28</td>
<td>1050</td>
<td>152.85</td>
</tr>
<tr>
<td></td>
<td>1945</td>
<td>153.51</td>
</tr>
<tr>
<td>29</td>
<td>1200</td>
<td>152.71</td>
</tr>
<tr>
<td></td>
<td>1945</td>
<td>153.61</td>
</tr>
<tr>
<td>30</td>
<td>1945</td>
<td>153.28</td>
</tr>
<tr>
<td>31</td>
<td>1845</td>
<td>152.98</td>
</tr>
<tr>
<td>Date 1979</td>
<td>Time</td>
<td>Observed</td>
</tr>
<tr>
<td>-----------</td>
<td>-------</td>
<td>----------</td>
</tr>
<tr>
<td>September</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>1945</td>
<td>153.18</td>
</tr>
<tr>
<td>2</td>
<td>1200</td>
<td>153.21</td>
</tr>
<tr>
<td></td>
<td>1945</td>
<td>153.84</td>
</tr>
<tr>
<td>3</td>
<td>1245</td>
<td>153.68</td>
</tr>
<tr>
<td>4</td>
<td>2000</td>
<td>153.08</td>
</tr>
<tr>
<td>5</td>
<td>1000</td>
<td>153.43</td>
</tr>
<tr>
<td></td>
<td>1915</td>
<td>153.25</td>
</tr>
<tr>
<td>6</td>
<td>1945</td>
<td>152.48</td>
</tr>
<tr>
<td>7</td>
<td>1945</td>
<td>152.88</td>
</tr>
<tr>
<td>8</td>
<td>1945</td>
<td>154.05</td>
</tr>
<tr>
<td>9</td>
<td>1845</td>
<td>152.88</td>
</tr>
<tr>
<td>10</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>11</td>
<td>1300</td>
<td>152.18</td>
</tr>
<tr>
<td></td>
<td>1655</td>
<td>152.48</td>
</tr>
<tr>
<td>12</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>13</td>
<td>1910</td>
<td>153.03</td>
</tr>
<tr>
<td>14</td>
<td>1845</td>
<td>153.28</td>
</tr>
<tr>
<td>15</td>
<td>1100</td>
<td>152.73</td>
</tr>
<tr>
<td></td>
<td>2155</td>
<td>153.18</td>
</tr>
<tr>
<td>16</td>
<td>1240</td>
<td>152.68</td>
</tr>
<tr>
<td></td>
<td>2030</td>
<td>152.98</td>
</tr>
<tr>
<td>17</td>
<td>0930</td>
<td>152.48</td>
</tr>
<tr>
<td></td>
<td>2030</td>
<td>153.2</td>
</tr>
<tr>
<td>18</td>
<td>1955</td>
<td>153.26</td>
</tr>
<tr>
<td>19</td>
<td>0945</td>
<td>152.68</td>
</tr>
<tr>
<td>November</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>1700</td>
<td>150.71</td>
</tr>
<tr>
<td>December</td>
<td></td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>1505</td>
<td>150.72</td>
</tr>
</tbody>
</table>
APPENDIX IV

The cross sectional area $A$ of the Meager Creek station on December 10, 1979 was calculated to be $2.7 \, m^2$. To calculate the stream velocity the Manning equation is used which states that

$$u = \frac{1.49 \, R^{2/3} \, S^{1/2}}{n}$$

where $u$ is the velocity, $R$ is the hydraulic radius (ratio of cross sectional area to the wetted perimeter), $S$ is the gradient slope and $n$ is the Manning fraction coefficient. For the Meager Creek station, $R=0.76$, $S=2.63 \times 10^{-2}$ and a reasonable estimate of $n$ is 0.04 to 0.05. The Manning equation leads to $u = 1.23$ to $1.53 \, m/sec$. The discharge, $Q$, obtained from $Q=UA$, is therefore 3.28 to 4.10 m$^3$/s or approximately 3.7 m$^3$/s.
APPENDIX V

DECEMBER DAILY DISCHARGES
OF THE LILLOOET RIVER NEAR PEMBERTON (1977-1979)
December Daily Discharge of the Lillooet River, Near Pemberton in m/s (1977-1979)

<table>
<thead>
<tr>
<th>Day</th>
<th>1977</th>
<th>1978</th>
<th>1979</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>32.0</td>
<td>22.7</td>
<td>24.1</td>
</tr>
<tr>
<td>2</td>
<td>31.4</td>
<td>22.4</td>
<td>23.8</td>
</tr>
<tr>
<td>3</td>
<td>32.6</td>
<td>22.1</td>
<td>23.5</td>
</tr>
<tr>
<td>4</td>
<td>30.3</td>
<td>21.8</td>
<td>23.2</td>
</tr>
<tr>
<td>5</td>
<td>29.2</td>
<td>21.5</td>
<td>22.9</td>
</tr>
<tr>
<td>6</td>
<td>28.3</td>
<td>21.2</td>
<td>22.9</td>
</tr>
<tr>
<td>7</td>
<td>28.0</td>
<td>21.0</td>
<td>23.1</td>
</tr>
<tr>
<td>8</td>
<td>27.5</td>
<td>20.7</td>
<td>23.2</td>
</tr>
<tr>
<td>9</td>
<td>26.9</td>
<td>20.4</td>
<td>24.9</td>
</tr>
<tr>
<td>10</td>
<td>27.5</td>
<td>20.1</td>
<td>25.5</td>
</tr>
<tr>
<td>11</td>
<td>28.3</td>
<td>19.8</td>
<td>23.8</td>
</tr>
<tr>
<td>12</td>
<td>34.0</td>
<td>19.7</td>
<td>22.9</td>
</tr>
<tr>
<td>13</td>
<td>33.1</td>
<td>19.5</td>
<td>19.8</td>
</tr>
<tr>
<td>14</td>
<td>31.1</td>
<td>19.4</td>
<td>21.5</td>
</tr>
<tr>
<td>15</td>
<td>28.3</td>
<td>19.3</td>
<td>20.7</td>
</tr>
<tr>
<td>16</td>
<td>26.9</td>
<td>19.1</td>
<td>20.1</td>
</tr>
<tr>
<td>17</td>
<td>25.5</td>
<td>19.0</td>
<td>21.0</td>
</tr>
<tr>
<td>18</td>
<td>27.8</td>
<td>18.8</td>
<td>24.6</td>
</tr>
<tr>
<td>19</td>
<td>27.2</td>
<td>18.7</td>
<td>36.0</td>
</tr>
<tr>
<td>20</td>
<td>26.9</td>
<td>18.5</td>
<td>32.6</td>
</tr>
<tr>
<td>21</td>
<td>26.6</td>
<td>18.4</td>
<td>29.7</td>
</tr>
<tr>
<td>22</td>
<td>26.3</td>
<td>18.3</td>
<td>28.0</td>
</tr>
<tr>
<td>23</td>
<td>26.1</td>
<td>18.1</td>
<td>26.6</td>
</tr>
<tr>
<td>24</td>
<td>25.8</td>
<td>18.0</td>
<td>25.5</td>
</tr>
<tr>
<td>25</td>
<td>25.5</td>
<td>17.8</td>
<td>24.6</td>
</tr>
<tr>
<td>26</td>
<td>24.9</td>
<td>17.7</td>
<td>24.4</td>
</tr>
<tr>
<td>27</td>
<td>24.6</td>
<td>17.6</td>
<td>23.5</td>
</tr>
<tr>
<td>28</td>
<td>24.4</td>
<td>17.4</td>
<td>22.9</td>
</tr>
<tr>
<td>29</td>
<td>24.1</td>
<td>17.3</td>
<td>22.4</td>
</tr>
<tr>
<td>30</td>
<td>23.8</td>
<td>17.1</td>
<td>22.1</td>
</tr>
<tr>
<td>31</td>
<td>23.8</td>
<td>17.0</td>
<td>21.9</td>
</tr>
<tr>
<td>Total</td>
<td>858.7</td>
<td>600.4</td>
<td>751.7</td>
</tr>
<tr>
<td>MEAN</td>
<td>27.7</td>
<td>19.4</td>
<td>24.2</td>
</tr>
<tr>
<td>MAX</td>
<td>34.0</td>
<td>22.7</td>
<td>36.0</td>
</tr>
<tr>
<td>MIN</td>
<td>23.8</td>
<td>17.0</td>
<td>19.8</td>
</tr>
</tbody>
</table>
References


Hubbert, M.K., The theory of groundwater motion, J. Geology, 48(8), Part 1, 785-944, 1940.


