APPLICATION OF GEOPHYSICS AND NUMERICAL MODELLING IN STUDYING AQUIFER HETEROGENEITY AND NITRATE TRANSPORT, ABBOTSFORD-SUMAS AQUIFER, BRITISH COLUMBIA, CANADA AND WASHINGTON, USA

by

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ABSTRACT

Heterogeneity within the sand and gravel deposits of the Abbotsford-Sumas aquifer has a significant impact upon groundwater movement and nitrate transport. Using GPR and borehole logging, the scale of heterogeneity was determined, with fining upward sequences up to 5 m thick and continuous over 10's of metres. Smaller heterogeneities were also identified visually in a local gravel pit. Various approaches were examined to represent this heterogeneity within a local groundwater flow model. The use of vertical anisotropy proved to be most realistic. Ages determined from the model were 60-80% lower than measured isotopic ages due to the inability to adequately represent the tortuosity of the flow paths. The spatial distribution and temporal variation of nitrate in the aquifer provided initial and calibration nitrate concentrations for the nitrate transport model. Nitrate concentrations thought to be reaching the aquifer based on recent BMPs are not sufficient to produce the observed nitrate concentrations.

Keywords: Abbotsford-Sumas aquifer; heterogeneity; ground-penetrating radar; borehole logging; nitrate transport modelling
EXECUTIVE SUMMARY

Aquifer heterogeneity plays an important role in groundwater movement and, consequently, contaminant transport. The Abbotsford-Sumas aquifer in the Fraser Valley of British Columbia and Washington is a heterogeneous aquifer consisting of sand and gravel glacial outwash deposits where NO₃ contamination has been of ongoing concern for many years.

To assist in better understanding this problem, a dataset of all available NO₃ data from the aquifer was assembled. The temporal and spatial variability of these NO₃ data were examined and a data file was created as the calibration dataset for NO₃ transport modelling.

Most NO₃ concentration averages calculated for different time intervals and at all depths shallower than 50 m are consistently at or slightly above the maximum allowable concentration (MAC) of 10 mg N/L. There is an annual cyclic pattern in the NO₃ data that matches closely the groundwater levels recorded in the aquifer, suggesting a relationship to precipitation/recharge.

The NO₃ concentrations appear to be elevated along the southern portion of the study area, with the exception of the area directly south of the Abbotsford airport. The lack of fertilizer application at the airport combined with a southerly flow of groundwater may account for this.

Ground penetrating radar (GPR) and borehole geophysical logging are used to investigate the scale of aquifer heterogeneity, and to determine what potential impact this could have on groundwater flow and NO₃ transport at a local scale. At this scale the glaciofluvial deposits comprising the aquifer are complex, with interbedded and cross-cutting sands and gravels. These complex permeable pathways might variably influence the movement of NO₃ through the aquifer.

The geophysics investigation was conducted at Agriculture and Agri-Food Canada's Pacific Agri-Food Research Centre (PARC) in Abbotsford. The borehole logs suggest that there
are fining-upward sequences (within coarse sediments) on the scale of 3 to 10 m and these sequences may extend laterally for 10 m or more. The GPR profiles suggest that some reflectors may be continuous over 100's of metres. At a site scale, the lateral continuity of these layers may be significant as they may contribute to permeable pathways for NO₃ migration. However, at a regional scale, it is unlikely that these sequences are laterally continuous, therefore, for transport modelling, it is unlikely that the heterogeneity could be represented.

Various approaches were taken to represent the observed heterogeneity within a 20 km² subset model of the regional groundwater flow model, which encompasses the PARC site. Water level contours and flow paths were used as boundary conditions along the edges of the model.

Initial attempts to refine the lithology within the model by re-examining lithology logs were unsuccessful as these logs did not provide sufficient data. In addition, the relationship between the natural-gamma count and grain size analysis data from several of the boreholes at the PARC site was examined. The large sample interval (1.2 m) of the grain size analysis data was too crude to identify any small scale relationship between these two datasets.

Another approach was based on the scale of heterogeneity observed within the geophysical data. The 5 m repeating fining upwards sequences were divided into 5 equally spaced layers with decreasing K values upward through the layers. A small test model showed that this increases the travel time, and distorts the travel paths of model particles. In the local scale model, it was unrealistic for these layers to continue beyond the 10's of metres indicated in the geophysics.

Ultimately, the most realistic representation of the heterogeneity was to use vertical anisotropy. A sensitivity analysis showed that lowering Kₓ by 1 order of magnitude, or bringing it closer to Kᵧ had little effect on the model. Larger decreases in Kₓ greatly effect the calibration.
For the scenarios examined in the sensitivity analysis, particle tracking was completed for piezometers with isotopic ages. The model ages were generally 60-80% of the isotopic ages. Based on both the sensitivity analysis and the age comparison, the most realistic representation of the groundwater movement was the scenario with $K_z$ lowered by one order of magnitude.

NO$_3$ transport modelling was completed within the local groundwater flow model. Using modelled NO$_3$ concentrations reaching the water table and the agricultural land use, constant concentration zones were designated at the water table to represent mean annual conditions. The calibration data and initial NO$_3$ concentrations were extracted from the NO$_3$ distribution maps.

The first model scenario was setup with uniform zero initial NO$_3$ concentrations to investigate NO$_3$ transport within the aquifer. The NO$_3$ initially moves down into the aquifer, and then follows the general groundwater flow direction from north-west to south-east. Equilibrium concentrations are generally attained within only a few years at most observation points.

A second scenario was modelled to investigate whether or not a change in Beneficial Management Practices (BMPs) in the mid-1990s could be expected to have an impact on aquifer nitrate concentrations. The simulation used initial NO$_3$ concentrations based on the 1992-1994 NO$_3$ distribution. Nitrate transport was simulated over a 10 year period, over which time the concentrations were observed to decrease in the model. The model was poorly calibrated as observed concentrations are considerably higher than those predicted by the model. This suggests that either the source of nitrate is too low and that other sources should be considered and/or that the current concentrations reflect a much longer history of contamination.

The third scenario used initial NO$_3$ concentrations based on the 2002-2004 NO$_3$ distribution to examine what would happen to NO$_3$ concentrations within the aquifer if the current input concentrations were to continue for another 10 years. The NO$_3$ concentrations again decrease to equilibrium concentrations within a few years. However, the outcomes are thought not to be realistic given the poor calibration for Scenario 2.
For my father
who would have been so proud
And for my mother
who is
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Permission has been obtained from Diana Allen to reproduce the copyrighted report in Appendix I. This report is titled “Abbotsford-Sumas Aquifer – Compilation of a Groundwater Chemistry Database with Analysis of Temporal Variations and Spatial Distributions of Nitrate Contamination”.

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1 INTRODUCTION

1.1 Aquifer Heterogeneity

Aquifer heterogeneity plays an important role in groundwater movement and, consequently, contaminant transport. Whereas homogeneous aquifers are characterized by spatially uniform geologic and, therefore, hydraulic properties (hydraulic conductivity, K, and storage coefficient, S), heterogeneous aquifers can have considerable spatially-variable hydrogeologic properties. For example, in sedimentary aquifers (either unconsolidated or consolidated), heterogeneity is often the result of changes in depositional environment resulting in variations in unit thicknesses, lithologic facies changes, or the nature of layering (Fetter, 2001). In fractured aquifers, the geometry and intensity of fracturing can lead to considerable heterogeneity (Mackie, 2002).

Heterogeneity also occurs over a range of scales; from changes in grain size within a single bed, to juxtaposition of different rock types at a regional scale. Jackson (2003) explained the importance of field scale studies over lab tests in heterogeneous aquifers, where variations may be on a scale greater than the lab sample size. Lesmes et al. (2002) and Sharp et al. (2003) explored scale effects related to aquifer heterogeneity and how these relate to the original depositional environment of the fluvial system.

At some scale, however, it is reasonable to assume that the properties are relatively homogenous. Thus, when considering groundwater flow and transport, it is necessary to identify the scale of the heterogeneity in addition to its character. How this heterogeneity is represented in numerical models is the topic of considerable ongoing research (for example, Fogg, 2006), as rarely can aquifers be considered truly homogeneous at all scales. Various approaches have been
taken to attempt to model aquifer heterogeneity. For example, Koltermann and Gorelick (1996) used structure-imitating, process-imitating, and descriptive methods to describe heterogeneity patterns, while Heinz et al. (2003) used outcrop analogues of glaciofluvial gravel deposits to assign varying hydraulic parameters.

The use of non-invasive geophysics has been shown to provide a quick and cost effective way to examine aquifer heterogeneity compared to drilling boreholes or excavating. Ground penetrating radar (GPR) has been used at many locations to improve aquifer characterization within glaciofluvial outwash deposits (Lunt et al., 2004; Close et al., 2004; Rea, 1996; Rea and Knight, 1998; Rea et al., 1994; Rea and Knight, 2000). Much of the work done using borehole geophysics for aquifer characterization has been to examine aquifer lithology and stratigraphic correlation, and water quality. Natural gamma logging is generally used for examining lithology (West, 2002; Norris, 1972; Baldwin and Miller, 1979; Dixon-Warren and Stohr, 2003), while conductivity logs provide information about the lithology and the water quality (Alger and Harrison, 1989; Keys, 1989).

The effects of heterogeneity on groundwater movement and contaminant transport can be complex. In cases where this heterogeneity occurs as geologic layering, the hydraulic conductivity has been observed to be greater along the bedding than perpendicular to them (Schafer, 1998), thus resulting in a net vertical anisotropy. The effects of heterogeneity on pumping tests have been observed to result in large K and S variances (e.g., Kollet and Zlotnik, 2005). With respect to solute transport, Dagan and Indelman (1999) showed that higher conductivity zones within a heterogeneous aquifer can create shortcuts that result in earlier breakthrough of the solute at monitoring points compared to transport in homogeneous material.

Of particular interest to this project is how heterogeneity within a regional aquifer affects groundwater movement and nitrate transport over the scale of several kilometres. Previous studies (e.g., Zheng and Gorelick, 2003) have indicated that preferential flow paths and barriers to
flow can have a significant impact on plume-scale transport. Where preferential flow paths act to channel and readily distribute solutes, barriers to flow act to limit or restrict the rate of solute movement. In the Abbotsford-Sumas aquifer, this combination of preferential flow paths and barriers has the potential to result in complex distributions of nitrate concentrations.

1.2 Regional Context

The Abbotsford-Sumas aquifer was selected as a case study site to investigate the effects of heterogeneity on groundwater flow and transport. The aquifer is situated in the central Fraser Valley of southern British Columbia and northern Washington, US (Figure 1.1). It is the largest unconfined aquifer in the region, approximately 161 km² (62 sq miles) in area, and is roughly bisected by the Canadian-US border. The aquifer spans uplands centred on Abbotsford, BC and drains into three river valleys (Fraser River to the north, the Sumas River to the east, and the Nooksak River to the south). It is highly productive, and provides water supply for nearly 10,000 people in the US (towns of Sumas, Lynden, Ferndale, Everson and scattered agricultural establishments) and 100,000 in Canada, mostly in City of Abbotsford, but also in township of Langley (Mitchell et al., 2000). The aquifer is known to reach depths of 70 m (Liebscher et al., 1992), and it is thickest in the northeast where glacial terminal moraine deposits are found. The deepest part of the aquifer system in this region is located near the US-Canada border, beneath the City of Abbotsford, BC and toward Lynden, WA, but the most productive areas are near Sumas, WA in southwest end of the Sumas Valley.

The coastal climate of the Abbotsford-Sumas region is humid and temperate, with significant rainfall (1000 to 2100 mm) over most of the year. Recharge to the aquifer is primarily from direct precipitation, mostly from October to May and ranges from 900 to 1100 mm/year (Scibek and Allen, 2005) following the regional precipitation gradient, which increases from south to north. According to Liebscher et al. (1992), groundwater generally flows from north to south, towards the US, with minimal vertical flow within the aquifer, except in the near surface
through direct recharge. However, recent modelling by Scibek (2005) has shown flow to be more complex and affected by local drainage and geology. Several small creeks drain the aquifer, and the baseflow of these streams is sustained in the summer months by groundwater discharge (Berg, 2006).

Figure 1.1: Location of the Abbotsford-Sumas aquifer.

The Abbotsford-Sumas aquifer system consists of several interconnected unconfined and confined aquifers, mostly in coarse grained sediments of glaciofluvial drift origin, which were deposited during the Sumas Stade (11,000 – 10,000 B.P.) of the Fraser Glaciation (Armstrong et al., 1965). The coarse grained glaciofluvial deposits are heterogeneous in nature, based on the
complex hydrostratigraphy modelled for the region (Scibek and Allen, 2005). Layering present within these deposits may have a substantial impact on groundwater movement and nitrate transport within this aquifer. Coarser grained layers may act as preferential pathways to groundwater movement and contaminant transport.

Elevated concentrations of nitrates have been documented in the Abbotsford-Sumas aquifer since the early 1970's (Liebscher et al., 1992). Over the aquifer, the main source of nitrate is attributed to agricultural activities, specifically fertilizer application, associated with raspberry production (Liebscher et al., 1992; Zebarth et al., 1998). There are also a significant number of chicken farms present in the area, and the manure produced from these farms is another potential source of nitrate contamination. Recent efforts at reducing the levels of contamination were thought to have been unsuccessful, as evidenced by stable or increasing nitrate levels in many monitoring piezometers (Hii et al., 1999; Hii et al., in draft).

A regional groundwater flow model was previously developed for the Abbotsford-Sumas aquifer (Scibek and Allen, 2005) as part of a study funded by Environment Canada. This model was based on the geologic interpretations of 1000's of borehole lithology logs, as well as hydrogeologic information available for the aquifer. The model was developed to facilitate an investigation into the potential transport of nitrate in the aquifer, which is the focus of ongoing research at Simon Fraser University. One of the possible limitations of the regional model is that it may not adequately represent the aquifer heterogeneity at the scale necessary for accurately modelling nitrate transport. Regardless, there are considerable uncertainties in the actual scale of the heterogeneity present in the Abbotsford-Sumas aquifer and, indeed, how heterogeneity might be incorporated into a numerical model that is used to simulate groundwater flow and transport.
1.3 Purpose and Objectives of Research

The main purpose of this research project is to investigate the effects of heterogeneity on the transport of nitrate contamination through the Abbotsford-Sumas aquifer. The key objectives are:

1. to determine the spatial distribution and temporal trends of nitrate contamination within the Abbotsford-Sumas aquifer and look for evidence of the effects of heterogeneity on nitrate transport,

2. to assemble a calibration dataset (nitrate concentrations) for numerical modelling of nitrate transport,

3. to determine the nature and scale of heterogeneity present at the site scale using a combination of geophysical methods and geological observations of gravel pit face exposures,

4. to investigate potential methods to represent site scale heterogeneity in a local scale numerical model, and

5. to investigate and attempt to quantify the effect of aquifer heterogeneity on the transport of nitrate through the aquifer.

1.4 Scope of Work

To meet the objectives of this research, a four phase project was undertaken. Phase 1 involved compiling all existing groundwater chemistry data from the Canadian portion of the aquifer\(^1\) and examining the temporal and spatial distribution of nitrate in the Abbotsford-Sumas aquifer.

Phase 2 consisted of conducting ground-penetrating radar (GPR) and borehole geophysical surveys to assess the heterogeneity present within the aquifer. These surveys were

\(^{1}\)US chemical data were not available despite efforts to acquire these data.
conducted at the Pacific Agriculture Research Centre (PARC), located southeast of the Abbotsford International Airport. The surveys were complimented by geological ground-truthing at a gravel pit situated 200 m south of the PARC facility that is thought to be representative of the local geology.

In Phase 3, a local scale groundwater flow model was developed from an existing regional groundwater flow model. The regional and local model boundaries are shown in Figure 1.1. Using this model, various approaches were attempted to represent the heterogeneity observed in Phase 2, and to study the effects of the aquifer heterogeneity on groundwater flow. Groundwater ages, determined from particle tracking with the model, were compared with isotopic ages obtained by Wassenaar et al. (2006).

In Phase 4, the numerical model in Phase 3 that best fit the heterogeneity and the groundwater ages was used to simulate nitrate transport through the aquifer. The movement of nitrate in the aquifer was examined. The model was used to investigate whether or not a change in Beneficial Management Practices (BMPs) in the mid-1990s could be expected to have an impact on aquifer nitrate concentrations.

1.5 Study Site

The large extent of the Abbotsford-Sumas aquifer precluded a region-wide investigation of the effects of heterogeneity on nitrate transport; therefore, a smaller sub-area was selected for detailed investigation. The study site was at the Agriculture and Agri-Food Canada’s Pacific Agri-Food Research Centre (PARC) Abbotsford substation, located 2.5 km southeast of the Abbotsford International Airport (Figure 1.1). It is approximately 200 m by 400 m, and is used for agricultural research. The site was chosen due to site availability, existence of a number of piezometers on-site, and because there is increased nitrate contamination within the aquifer in this area (McArthur and Allen, 2005). Additionally, Environment Canada installed 10 piezometers at
this facility as part of its ongoing nitrate monitoring program in the aquifer. These piezometers provided an excellent control for both water table elevation and nitrate concentration data. The piezometers range in depth from 19.4 m to 46.4 m, with an average depth of 27.9 m. Lithologic data, collected when the piezometers were drilled, provided insight into the geologic framework beneath the site, although detailed grain size analyses were lacking.

The local scale numerical groundwater flow model was positioned around the PARC site. This 20 km² model was based on flow boundaries as observed in the regional groundwater model (Scibek and Allen, 2005).

The PARC substation is located within the Abbotsford outwash deposit of the Sumas Drift. This coarse sand and gravel unit was deposited during the Sumas Stade (11,000 – 10,000 BP) of the Fraser Glaciation (Armstrong et al., 1965).

An abandoned gravel pit, located 200 m to the south of the PARC substation (Figure 1.1), was also visited to investigate the local geological framework. A series of photographs were taken in a north-south direction along a wall within the gravel pit.

1.6 Thesis Outline

This thesis is comprised of a summary of a scientific report (full report in Appendix I), one journal article that has been submitted for publication in a peer-reviewed scientific journal, and two standard chapters. Some text and figures may be repeated in the various chapters as the chapters were intended to be published separately.

Chapter 2 summarizes and updates the results of a scientific report entitled “Abbotsford-Sumas Aquifer: Compilation of a Groundwater Chemistry Database with Analysis of Temporal Variations and Spatial Distributions of Nitrate Contamination,” (McArthur and Allen, 2005), which was prepared for the BC Ministry of Water, Land and Air Protection (MWLAP) (now BC Ministry of Environment (MOE)).
Chapter 3 comprises the manuscript related to the geophysical investigation. The thesis appendices contain an overview of the geophysical methods employed and the entire geophysical dataset collected during the field program. The ground-penetrating radar (GPR) data are presented in Appendix II. The borehole geophysical data, including natural gamma and electrical conductivity logs, are contained in Appendix III.

Chapter 4, written in standard thesis format, describes the methodology and results of the local scale numerical modelling exercise undertaken to investigate the effects of heterogeneity on groundwater movement through the aquifer. A portion of this chapter has been submitted for publication in the Proceedings of the Joint Canadian Geotechnical Society-International Association of Hydrogeologists (Canadian National Chapter) Conference to be held in Vancouver, BC, October 2006 (Chesnaux et al., 2006). The paper is entitled “Investigating the Scale of Heterogeneity and Implications for Nitrate Transport, Abbotsford-Sumas Aquifer, BC and WA” (Chesnaux et al., 2006). The lithology logs that were examined in this chapter are found in Appendix IV, and the grain size analysis data are found in Appendix V.

Chapter 5, written in standard thesis format, describes the methodology and results of the numerical modelling exercise undertaken to investigate the transport of NO₃ through the aquifer. This exercise provides the basis for a larger regional NO₃ transport model that will be undertaken. Chapters 4 and 5 combined are intended for publication in a scientific journal.

Chapter 6 provides the conclusions of the thesis.
2 NITRATE CONTAMINATION

This chapter summarizes and updates the report “Abbotsford-Sumas Aquifer: Compilation of a Groundwater Chemistry Database with Analysis of Temporal Variations and Spatial Distributions of Nitrate Contamination” that was submitted to the BC Ministry of Water, Land and Air Protection (MWLAP) (now BC Ministry of Environment (MOE)) in February 2005. The complete report can be found in Appendix I.

2.1 Introduction

Nitrate contamination of groundwater is a serious concern in agriculture-rich areas, such as the lower Fraser Valley in southwestern British Columbia, Canada and northern Washington State, USA. The 161 km$^2$ (62 sq miles) trans-national Abbotsford-Sumas aquifer straddles the Canada - USA border (Figure 2.1). The aquifer system consists of several interconnected unconfined and confined aquifers, mostly in coarse grained sediments of glaciofluvial drift origin, and spans uplands and three river valleys (lowlands or floodplains). The aquifer is highly productive, and provides water supply for nearly 10,000 people in the USA (towns of Sumas, Lynden, Ferndale, Everson and scattered agricultural establishments) and 100,000 in Canada, mostly in City of Abbotsford, but also in township of Langley (Mitchell et al., 2000). Figure 2.2 shows the location of major production wells in the Canadian portion of the aquifer as well as major city centres.
In previous studies, Environment Canada reported on temporal trends in monitoring wells in selected areas of the aquifer, based largely on their monitoring network. However, over the past several decades a considerable number of wells have been sampled by other government agencies, most notably the BC MOE (formerly the MWLAP). The majority of these data are contained in the provincial Environmental Monitoring System (EMS) database, with the rest in separate databases.
As part of this research project, a comprehensive nitrate (and other chemical species, where available) database was created from all of the existing groundwater chemistry on the Canadian side of the Abbotsford-Sumas aquifer up to April 2004. Despite considerable effort to locate US data, none was forthcoming from the relevant US government agencies. Temporal and spatial trends in the nitrate data were analyzed in order to identify the long- and short-term variations in concentrations of nitrate, and relate these to well depth and potential driving forces, such as precipitation. These data will be used further as calibration data for nitrate transport modelling of the aquifer.

Figure 2.2: Central Fraser Valley location map showing the Canadian portion of the Abbotsford-Sumas aquifer examined (in green), major Canadian production wells, along with cities and towns, topography, international boundary, and major rivers.
2.2 Background

Nitrates enter groundwater from either natural or anthropogenic sources. Sources of natural nitrates include soil nitrogen, nitrogen-rich geologic deposits and nitrogen from the atmosphere. Sources of anthropogenic nitrate include fertilizers, farm animal manure, septic tank drainage, and leaching of the soil due to irrigation (Madison and Brunett, 1985). There has been some suggestion that ammonia emissions from poultry farms may contribute to nitrate loading (Sutherland, pers. comm., 2006). Nitrate in the soil is affected by the nitrogen cycle, which includes the following five stages (Canter, 1997):

1. Nitrogen fixation: the conversion of gaseous nitrogen into ammonium by bacterial conversion for use by plants and animals.

2. Ammonification: the conversion of organic nitrogen into ammonia during the decomposition of organic matter.

3. Assimilation: a biochemical process that uses ammonium or nitrate to form nitrogen-containing compounds.


5. Denitrification: the conversion of nitrate into nitrite, nitrous oxide, nitrogen dioxide and nitrogen gas by anaerobic bacteria.

Of these five stages, only denitrification results in significant, long-term decreases in nitrate concentrations because the products are unlikely to convert back to nitrate (Wassenaar, 1995). Adsorption and anion exchange may contribute to decreases in NO₃⁻, but these processes are thought to be rare due to the limited clay component of the aquifer material.

Elevated concentrations of nitrates have been documented in the Abbotsford-Sumas aquifer since the early 1970's (Liebscher et al., 1992). Over the aquifer, the main source of nitrate
is attributed to agricultural activities (Liebscher et al., 1992), with raspberries being the predominant crop above the Canadian portion of the aquifer (Hii et al., 1999). Fertilizer application practices that are associated with raspberry production have been identified as significant contributors to nitrates in the aquifer (Liebscher et al., 1992; Zebarth et al., 1998; Mitchell et al., 2003). There are also a significant number of chicken farms present in the area, and the manure produced from these farms is another potential source of nitrate contamination.

A study of nitrogen origin and fate in the aquifer (Wassenaar, 1995) using nitrogen and oxygen analysis, indicated that soil nitrate was predominantly derived from nitrification, manure, and, to a lesser extent, from ammonium-based fertilizers. However, recent work by Wassenaar (2006) may suggest a shift in nitrogen sources, away from manure sources towards inorganic fertilizer sources. This shift in source may follow a recent (since 1990) shift in agricultural practice away from the use of manure fertilizer to synthetic fertilizer as a response to a challenge to the industry to reduce nitrate loading (Sutherland, BC Ministry of Agriculture, personal communication). Hii et al. (in draft) indicate that the industry as a whole responded to the challenge; however, the results of the latest survey (Hii et al., in draft) indicate that the extent of nitrate contamination throughout the aquifer has not changed dramatically.

2.3 The Dataset

The data used for this project were compiled from several different sources. Data were provided by Dennis Barlow and Marc Zubel of BC MOE and accessed directly from the MOE EMS (BCMWLAP, 2004), with information dating back to 1979. Environment Canada (EC) also provided data back to 1971. The EC data, along with a majority of MOE data, were collected by these government agencies. There also exist data of opportunity, which were collected from residential wells and taps. These data of opportunity are likely to have a higher uncertainty as there is limited control over the sampling techniques used (water sample taken from residential tap). The government collected data are expected to be less affected by sampling errors, and
more representative of depth-specific concentrations (piezometers used). All samples were
analyzed in accredited government laboratories using standard practices.

For many wells in the EMS database, UTM coordinates and Well Tag Numbers (WTN)
were not available. Many of the locations were found only through a careful search of the EMS
website, locational data provided by EC, and well location information compiled in a report by
Piteau Associates (Dakin, 2003). However, many wells were never located.

The data were reorganized into a single dataset (see Appendix I-A), which includes
information about well locations and well depth, along with all of the available groundwater
chemistry data. Nitrate data from the dataset were extracted to analyze spatial and temporal
trends, as discussed herein, and to create a data file of observed data for calibration of a nitrate
transport model (two models were developed; one local one as part of this research, and a larger
scale model as part of ongoing research). This reduced data file is found in Appendix I-C. The
observations include the name of the well, the location of the sample point, the time of
measurement, and the concentration of nitrate observed.

2.4 Temporal Variations

2.4.1 Long Term Trends

As a first step to understanding the nature of nitrate contamination within the Abbotsford-
Sumas aquifer, the temporal variations of the nitrate concentrations were examined. Long term
monitoring of many of the wells began in the 1980’s, with a few wells going back as far as 1970.
For this project, long term monitoring wells include those that were sampled over a minimum of
three years. Wells without long term monitoring were excluded from further temporal analysis.

Figure 2.3 shows a plot of all of the nitrate concentration data from wells with long term
nitrate concentration monitoring (between 1971 and 2004). The anomalously high nitrate
concentrations are from EC piezometer ABB3. The positive, but small, linear trend supports the
observation that the NO$_3$-N concentrations have been increasing over time (Hii et al., 1999; Liebscher et al., 1992), although a lack of early data may accentuate to the increasing trend. Also shown on Figure 2.3 are the 49 sample (pink) and 99 sample (blue) moving averages. The average concentration over this time period is 13±9 mg N/L. For comparison purposes, Fetter (1993) indicated that background nitrate concentrations in groundwater are typically around 3 mg N/L. The average nitrate concentration measured in the Abbotsford-Sumas aquifer exceeds the maximum allowable concentration (MAC) guideline as set by Health Canada (2005) for NO$_3$-N in drinking water of 10 mg N/L (shown as a red line in Figure 2.3).

Figure 2.4 shows the latter portion of the dataset (1998 to 2004), including the 49 sample and 99 sample moving averages. The average concentration over this time period is 14±9 mg N/L, which is similar to the results from the complete dataset. Note that the trend in nitrate concentration is negative over this time period, suggesting somewhat improved conditions in the aquifer in recent years, which could be associated with a decrease in recharge and/or reduced nitrate loading due to compliance with the BMPs. In Figure 2.5, the period of plotting is further reduced to 2002 - 2004. The negative trend in nitrate concentration in this plot is more significant than over the larger time interval. These somewhat conflicting results suggest that there are variable trends depending on the start date, and that there may be some climatic variation in the nitrate concentration trends.
Figure 2.3: Nitrate data from wells with long term monitoring in the Abbotsford-Sumas aquifer. Entire historic dataset is shown (1970 - 2004).
Figure 2.4: Nitrate data from wells showing a portion of the long term dataset (1998 – 2004).
Figure 2.5: Nitrate data from wells showing a portion of the long term dataset (2002 – 2004).
2.4.2 *Seasonal Variations – 1998 to 2004*

Seasonal variations from 1998 to 2004 were examined. Figure 2.6 shows the 99 sample moving average, along with the water level variations recorded in provincial observation wells #2, #8 and #299. Mean monthly precipitation as recorded at the Abbotsford International Airport is also shown. The graph reveals an annual cyclical pattern in the nitrate concentrations across the entire aquifer; the rise and fall in nitrate concentration has the same general pattern as water levels recorded within the aquifer. This cycle appears to be linked to the annual precipitation (and recharge) variation, although both the water levels and nitrate concentrations trail precipitation cycles by a few months. These results are consistent with the hydraulic response time of the aquifer to precipitation events. Thus, it is speculated that increased rainfall over the late fall and winter months acts to mobilize nitrates, likely originating from agriculture, and transports the nitrate down to the water table through the thick vadose zone.

2.4.3 *Temporal Variations - Depth Profiles*

Wells were grouped based on the depth of the water samples. The depth that was used was either the centre of the screen interval, if this information was provided, or the depth of the well, when no screen information was available, as most wells are anticipated to be screened near the bottom of the well. The average nitrate concentration and the standard deviation at each depth interval were calculated and are reported in Table 2-1.
Figure 2.6: Nitrate data from wells showing a portion of the long term dataset (1998 - 2004), the hydrographs for observation wells #2, #8 and #299 (BC MOE, 2006), and mean monthly precipitation as recorded at the Abbotsford International Airport (Environment Canada, 2004).
Table 2.1: Average nitrate values at various depth ranges in the Abbotsford-Sumas aquifer.

<table>
<thead>
<tr>
<th>Depth Range (m)</th>
<th>Average ( \text{NO}_3 ) Concentration (mg N/L)</th>
<th>Standard Deviation (mg N/L)</th>
<th>Number of Samples</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-10</td>
<td>13</td>
<td>7</td>
<td>1471</td>
</tr>
<tr>
<td>10-20</td>
<td>16</td>
<td>13</td>
<td>824</td>
</tr>
<tr>
<td>20-30</td>
<td>12</td>
<td>8</td>
<td>870</td>
</tr>
<tr>
<td>30-40</td>
<td>16</td>
<td>5</td>
<td>335</td>
</tr>
<tr>
<td>40-50</td>
<td>13</td>
<td>5</td>
<td>274</td>
</tr>
<tr>
<td>50+</td>
<td>7</td>
<td>8</td>
<td>57</td>
</tr>
</tbody>
</table>

The depth profile data, on average, show that there are more instances of concentrations above the MAC of 10 mg N/L than below it. Average nitrate concentrations are typically less than the MAC of 10 mg N/L in wells greater than 50 m depth. However, the limited number of wells greater than 50 m depth may bias these results. Average concentrations per depth range also do not consistently decrease with depth, as might be expected due to increased dilution due to mixing at depth, away from source areas. In fact, average concentrations are roughly similar for all depth ranges, except > 50 m. The most likely explanation for the presence of higher concentrations at depth is the presence of heterogeneity, which results in preferential pathways for nitrate transport. Water that infiltrates from source areas with high nitrate loading may have followed preferential flow paths, producing elevated concentrations at some depths and not at others. Hii (1999) similarly observed that nitrate concentrations do not systematically decrease or increase with depth at Environment Canada sites where wells of varying depths are monitored, although the general trend observed is a decrease in concentration with depth.

Within all depth intervals, except the 20 to 30 m interval, there is a slight increase in nitrate concentration with time (McArthur and Allen, 2005). In wells between 20 and 30 m depth, there is a decrease in the nitrate concentration with time.
2.5 Spatial Distributions

2.5.1 Spatial Distribution – Time Intervals

The spatial distribution of nitrate concentration across the Abbotsford-Sumas aquifer is of particular interest, as it may identify hotspots for contamination, and possibly provide insight into the origin of elevated nitrates if correlated with land use maps. Several distribution maps were created from the available data; these are shown in Figures 2.7 through 2.9. The data were examined over three time intervals (1992-1994; 1997-1999; 2002-2004). In all three time intervals, there are limited data available. This may result in contouring artefacts, particularly around the edges. However potential hotspots are highlighted, and these areas of elevated nitrate concentrations correlate with those of Hii et al. (1999).

All three time intervals show elevated nitrate concentrations in the part of the aquifer south and southeast of the Abbotsford airport. These results agree with Hii et al. (1999; in draft), which show a similar distribution of nitrate in the area south of the airport (from selected data only). The land use in this area is primarily agriculture, with most of the area covered by raspberry farms, followed by poultry farms (BC MAL, 2005). Fertilizers from raspberry farms and manure stock piles from poultry farms are two likely sources of NO₃ to the aquifer. Nitrate loading, associated with agricultural practices, does not occur on the Abbotsford airport property, which results in lower nitrate concentrations in the area directly south of the airport. This is consistent with groundwater flow directions, which are dominantly to the south (Liebscher et al., 1992). However, local variations in groundwater flow direction do occur, primarily near streams and lakes, and in areas where there are changes in aquifer hydraulic properties based on the results of numerical modelling (Scibek and Allen, 2005). Only the 2002-2004 interval provides information about the western part of the aquifer, and shows that the nitrate concentrations are generally not above 10 mg N/L.
2.5.2 Spatial Distribution - Changes Over Time

In order to determine if the locations of elevated nitrate concentrations are changing over time, the differences in concentrations over a 10 year period between the 1992-1994 and 2002-2004 were contoured and plotted (Figure 2.10). Only data that were common to both time periods were used to calculate the difference in concentration. As with the individual time periods, contouring artefacts may be present due to the limited number of data. Within the overlapping map areas nitrate concentrations appear to have both increased by up to 10 mg N/L, and decreased by 5 mg N/L. Higher concentrations are observed to the southeast of the airport, but lower concentrations appear to occur to the east. However, the dataset is sparse, and, consequently, there is ambiguity in these results. A comparison of the NO₃ concentrations between 1992-1994 and 2002-2004 is shown in Figure 2.11. The trend is for an increase in concentration over the 10 year period (note higher intercept), but many wells show a decrease in NO₃ concentration. Several wells show a significant increase in NO₃ concentration during this time period.
Figure 2.8: Nitrate distribution from 1997 to 1999 in the Abbotsford-Sumas aquifer.
Figure 2.9: Nitrate distribution from 2002 to 2004 in the Abbotsford-Sumas aquifer.

LEGEND
Source of Chemistry Data
• BCMWLAP
• EMS
• Env Can
• Matsqui

NO3-N Concentration (mg N/L)
- 0.16 - 2
- 2.01 - 4
- 4.01 - 6
- 6.01 - 8
- 8.01 - 10
- 10.01 - 12
- 12.01 - 14
- 14.01 - 16
- 16.01 - 18
- 18.01 - 20
- 20.01 - 36.4
Figure 2.10: Change in nitrate distribution between 1992-1994 and 2002-2004 in the Abbotsford-Sumas aquifer.
2.6 Discussion

The majority of the data show that historic nitrate values within the Canadian portion of the Abbotsford-Sumas aquifer are elevated above normal background concentrations of 3 mg N/L. Most of the averages calculated from long term data or seasonal data were above the MAC of 10 mg N/L, with concentrations in one well exceeding 90 mg N/L. These results are consistent with previous work on nitrate concentrations from selected wells (Hii et al., 1999), and the results of a more recent survey of nitrate concentrations in 2005 (Hii et al., in draft). Note that the data up to April 2004 from Hii et al. (in draft) were included in the current dataset. Hii et al.’s study also included data from late 2004 and a few samples from 2005. These studies combined suggest that elevated nitrate concentrations exist in this aquifer, and that despite efforts on the part of the agriculture industry to reduce nitrate loading since 1990, there has not been a significant reduction in nitrate concentrations. Nonetheless, the temporal analysis conducted as part of this
study suggests some improvement in groundwater quality, but that the fluctuations may be more driven by precipitation variation than by changes in land use.

The nitrate concentrations within the Abbotsford-Sumas aquifer follow an annual cyclic pattern, which correlates with water levels in the aquifer, and appears to relate to the timing of precipitation events, and also, likely, timing of fertilizer application. Higher precipitation results in greater infiltration (recharge) (Scibek and Allen, in press), and, consequently, there is a potential for higher amounts of nitrate leaching from the soils and entering the groundwater. The timing of fertilizer application in agricultural practices may also contribute to the observed fluctuations, as this is the main source of nitrate in the Abbotsford-Sumas aquifer. The role of irrigation is uncertain, but may be a factor in mobilizing nitrate (Sutherland, personal communication, 2006).

The nitrate concentrations do not show any significant variation with depth, but neither is there consistent decrease (or increase) in concentrations with depth. Below 50 m, the concentrations drop to concentrations generally below the MAC. The variations in the vertical distribution of nitrate are anticipated to be associated with the presence of permeable pathways due to the heterogeneity of the aquifer.

The spatial distributions of nitrate within the aquifer have changed somewhat in the 10 years between the 1992-1994 data and the 2002-2004 data. Variations in the range of -5 to +10 mg N/L have occurred over this time period, but the results are somewhat inconclusive due to the sparse dataset. Significant changes in the nitrate concentrations are not expected unless there is a significant change in fertilizer use and application practices, as agriculture continues to dominate the land use in this area.
3 RESOLVING AQUIFER HETEROGENEITY USING GROUND-PENETRATING RADAR AND BOREHOLE GEOPHYSICAL LOGGING WITH IMPLICATIONS FOR NITRATE TRANSPORT

To be submitted to the Journal of Environmental and Engineering Geophysics.

3.1 Abstract

Ground penetrating radar (GPR) and borehole geophysical logging are used to investigate the scale of heterogeneity within the trans-national Abbotsford-Sumas aquifer, situated in British Columbia (BC) and Washington State (WA). The aquifer provides a source of water for approximately 100,000 people, although nitrate contamination has become a significant problem over the last 30 years owing to intensive agricultural activities on both sides of the border. At a scale necessary for modelling nitrate transport, the glaciofluvial deposits comprising the aquifer are complex, with interbedded and cross-bedded sands and gravels. It is anticipated that local heterogeneity will result in complex permeable pathways, which might variably influence the movement of nitrate through the aquifer. A series of GPR surveys and borehole logs were collected at Agriculture and Agri-Food Canada's Pacific Agri-Food Research Centre (PARC) in Abbotsford, BC to investigate the scale of heterogeneity present locally, and to determine what potential impact this could have on groundwater flow and nitrate transport at a local scale. The borehole logs suggest that there are fining-upward sequences (within coarse sediments) on the scale of 3 to 10 m. Representing this vertical heterogeneity will be important in vadose zone transport modelling as there will a strong impact on the vertical anisotropy in hydraulic conductivity due to variably saturated conditions. The GPR profiles show that these sequences extend laterally a few meters between nested piezometers, and perhaps up to several 10s of metres. At a site scale, the lateral continuity of these layers may be significant as they may
contribute to permeable pathways for nitrate migration. However, at a larger scale, it is unlikely that these sequences are laterally continuous. Therefore, for transport modelling it is unlikely that the heterogeneity could be represented.

3.2 Introduction

This study uses a combination of borehole geophysical logging and ground penetrating radar (GPR) surveying, with geological constraints and interpretations, to facilitate the interpretation of aquifer architecture heterogeneity, which may influence nitrate transport at a local scale.

The study site is located within the Abbotsford-Sumas aquifer, which is situated in the central Fraser Valley of southern British Columbia, Canada and northern Washington, USA (Figure 3-1). The aquifer is the largest unconfined aquifer in the region, covering an area of approximately 161 km\(^2\) (62 sq. miles), and is roughly bisected by the Canada-USA border.

Elevated concentrations of nitrates have been documented in the Abbotsford-Sumas aquifer since the early 1970's (Liebscher et al., 1992). Over the aquifer, the main source of nitrate is attributed to agricultural activities, specifically fertilizer application associated with raspberry production (Liebscher et al., 1992; Zebarth et al., 1998). There are also a significant number of chicken farms present in the area, and the manure produced from these farms is another potential source of nitrate contamination. Recent efforts at reducing the levels of contamination were initially thought to have been unsuccessful, as evidenced by stable or increasing nitrate levels in many monitoring piezometers (McArthur and Allen, 2005; Hii et al., 1999); however, more recent data (Hii et al., in draft) suggests that nitrate levels are starting to improve.
To better understand the nature of the regional distribution of nitrate, the rate of movement of nitrate, and the relation of nitrate levels to current land use, nitrate transport modelling is being conducted based on an existing regional scale groundwater flow model (Scibek and Allen, 2005). While this model was constructed using lithologic data from over 5000 water wells and, thus, captures effectively the significant heterogeneity present at a regional scale, its ability to resolve nitrate pathways is limited due to the lack of detailed information on the
nature of heterogeneity at a local scale. The transport of nitrate within the aquifer is anticipated to follow tortuous pathways, as a direct consequence of the high degree of heterogeneity present over a range of scales.

To investigate heterogeneity at a local scale, several GPR surveys were conducted at Agriculture and Agri-Food Canada’s Pacific Agri-Food Research Centre (PARC) substation in Abbotsford, BC (Figures 3-1 and 3-2). Borehole geophysical logs, including natural-gamma and electromagnetic induction (conductivity), were acquired in nine piezometers around the perimeter of the site. These piezometers provide not only water table elevation data, but also nitrate concentrations at specific depths. Geologic constraints are provided from borehole lithologic (driller’s) logs as well as from an abandoned gravel pit (cross-sectional pit face) situated approximately 200 m south of the PARC site.

Figure 3.2: Site map showing the locations of the GPR lines and the nested piezometers. The black square on Line 1 indicates the location of the CMP survey midpoint.
3.3 Background

3.3.1 Hydrogeology of the Abbotsford-Sumas Aquifer

The Abbotsford-Sumas aquifer is located on a broad outwash plain, which is elevated above the adjacent river floodplains (Figure 3-1). The outwash terrace slopes southward, and terminates in escarpments along the Nooksack River floodplain. To the west, there is a drainage divide along hilly terrain of the township of Langley, BC. To the east is the Sumas River valley, a large sediment-filled bedrock valley. The uplands are centered on the City of Abbotsford, BC and extend westward through Langley, BC and south to Lynden, WA. The aquifer is highly productive, and provides water supply for nearly 10,000 people in the USA (towns of Sumas, Lynden, Ferndale, Everson and scattered agricultural establishments) and 100,000 in Canada, mostly in the City of Abbotsford, but also in the township of Langley (Mitchell et al., 2000). The coastal climate is humid and temperate, with 1000 to 2100 mm mean annual rainfall over most of the year. Recharge to the aquifer (900 to 1100 mm) is primarily from direct precipitation, mostly from October to May (Scibek and Allen, 2005; Scibek and Allen, in press).

Liebscher et al. (1992) determined that in the southern part of the aquifer, groundwater generally flows from north to south, with some local variations. There is minimal vertical flow within the aquifer, except in the near surface through direct recharge. This general flow direction was confirmed through numerical modelling (Scibek and Allen, 2005), although heterogeneity of the surficial sediments was observed to redirect groundwater locally, particularly around streams. A significant component of the groundwater flow was also found to occur eastward, toward the Sumas River (Figure 3-1). Groundwater discharge occurs through spring flow and seepage to small streams and rivers. The largest rivers, which are hydraulically connected to the aquifer system, are the Nooksack River and the Sumas River. These are almost exclusively discharge zones. Small streams on the uplands, and small lakes, have more complex and temporally-varying aquifer interactions. To the north is the Fraser River floodplain, where a small
component of groundwater discharge occurs just to the west of Sumas Mountain in BC. On a large scale, the entire valley is called Fraser River Valley, or, more specifically, the central Fraser Valley (also called Fraser Lowland).

The aquifer is comprised of coarse-grained sediments of glaciofluvial drift origin, primarily uncompacted sands and gravels of the Sumas Drift, with lenses of sand, silt, and clay (Armstrong et al., 1965). The Abbotsford Outwash, the deposit within the Sumas Drift beneath the PARC site, is a glacial outwash deposit, which was deposited during the Sumas Stade (11,000 – 10,000 B.P.) of the Fraser Glaciation. There is significant heterogeneity of the hydrostratigraphic units as evidenced by the lithologies encountered and logged during drilling. This heterogeneity can be expected to result in complex groundwater paths at both regional and local scales. The aquifer is known to reach depths of 70 m (Liebscher et al., 1992), and it is thickest in the northeast where glacial terminal moraine deposits are found. The deepest part of the aquifer system in this region is located along the Canada-USA border, beneath the City of Abbotsford, BC and toward Lynden, WA, but the most productive areas are near Sumas, WA in southwest end of the Sumas Valley.

The Abbotsford-Sumas aquifer is mostly unconfined in the outwash plains, but significant confined pockets are present. For example, a large part of the aquifer within Sumas Valley is confined. Smaller confined aquifers are also located to the northwest, in the uplands; however, these aquifers are poorly connected to the Abbotsford-Sumas aquifer (Scibek and Allen, 2005). Laterally, the valley sediments are confined by the Tertiary bedrock surface, which outcrops as mountains on both sides of Sumas Valley, and as small outcrops south of Nooksack River. The aquifer is underlain by an extensive glaciomarine deposit, the Fort Langley Formation, which outcrops in the uplands to the west. The distribution of unconfined and confined portions of the aquifer will have a bearing on the distribution of nitrate contamination as confining units tend to protect underlying units from surface contamination.
3.3.2 **Geophysical Approach**

A quick and cost effective way to examine aquifer heterogeneity is through the use of surface geophysics. GPR has been used to improve aquifer characterization at a number of sites. For example, Lunt et al. (2004) used GPR at the Sagavanirktok River deposits in Alaska, and were able to identify numerous features indicative of these Quaternary glaciofluvial outwash deposits. Close et al. (2004) examined the presence of preferential flow paths at two locations in New Zealand, which contained heterogeneous aquifers. Preferential pathways were identified within these aquifers by correlating the GPR results to core logging.

Within the Fraser Valley, an extensive GPR survey was previously conducted in the nearby Brookswood aquifer in BC, which is a similarly heterogeneous aquifer comprised of Sumas Drift (Rea, 1996; Rea and Knight, 1998; Rea et al., 1994; Rea and Knight, 2000). Aquifer characterization was conducted by identifying and mapping radar architectural elements (Rea, 1996; Rea and Knight, 2000). These architectural elements were identified as hydrogeological units and were combined with drillers' logs to reconstruct the depositional environment. The depth of penetration on the GPR profiles ranged from about 5 m in the eastern part of the aquifer to about 15 m in the western part. This variation is due to changes in attenuation of the GPR signal in different locations. The highly heterogeneous nature of the Brookswood aquifer was apparent in the GPR data, especially the presence of electrically conductive clay bodies, which were suggested to be no-flow boundaries. As part of that study, a single 20 m long GPR line was also collected at the PARC site, which showed that there were several nearly horizontal and slightly dipping reflectors visible in the top 6m of the profile.

Much of the work done using borehole geophysics for aquifer characterization is to examine aquifer lithology and stratigraphic correlation, and water quality. Of the various tools available, aquifer lithology is commonly examined using both natural-gamma and conductivity logging (Cromwell, 1992; Nobes and Schneider, 1996; Barrash and Morin, 1997; Pullan et al.,
2002; Siron and Segall, 1997; Paillet and Reese, 2000). Because natural-gamma is sensitive to clay content and grain size, it is often the only geophysical log used to examine aquifer lithology (West 2002; Norris, 1972; Baldwin and Miller, 1979; Dixon-Warren and Stohr, 2003). Since the conductivity probe measures electrical conductivity of both the material as well as the groundwater, conductivity logs can also be used to examine groundwater quality by comparison to natural-gamma and geologic logs (Alger and Harrison, 1989; Keys, 1989).

In 1993, as part of the Fraser Lowland Hydrogeology Project, two piezometers at the PARC site (92-3 and 92-4) were logged using natural-gamma, conductivity and magnetic susceptibility sondes (GSC, 2003). These data are used for comparison with our logs. Irving and Knight (2003) conducted cross-hole GPR logging between different pairs of piezometers at the PARC site in order to investigate saturation-dependent radar wave velocity anisotropy in both the saturated and unsaturated zone. Their findings indicate that in the vadose zone, significant vertical anisotropy can result due to the strong dependence of dielectric constant on saturation and the pronounced saturation heterogeneity that can exist. As the overall saturation in the vadose zone decreases, fine grained layers preferentially retain water, while coarse-grained layers preferentially drain; this process enhances the dielectric contrast between layers (Irving and Knight, 2003). They determined an average vertical radar velocity of 0.12 m/ns in the vadose zone. Their survey was conducted in late July/early August when ground saturation levels are near their lowest in the aquifer (Irving, personal communication, 2006). They also completed additional cross-hole GPR logs at the site in 2004, which are still being analyzed (Irving, personal communication, 2006). Finally, Pullan et al. (2000) completed a single shallow seismic-reflection survey line along Boundary Road overlying the Abbotsford-Sumas aquifer. The quality of the data was considered to be poor due primarily to dry coarse-grained surface materials.
3.4 Current Investigation

3.4.1 Site Description

All of the geophysical surveying for this current investigation was completed at the PARC site in Abbotsford, BC. The PARC site is approximately 200 m by 400 m, and is used for test crops. As part of its monitoring program, Environment Canada installed 10 piezometers at this facility (Figure 3-2); these piezometers range in depth from 19.4 m to 46.4 m, with an average depth of 27.9 m. The site was chosen due to site availability, existence of a number of piezometers on-site, and because there is increased nitrate contamination within the aquifer in this area (McArthur and Allen, 2005). The field work was completed over 5 days during May and August of 2005. Water levels in the piezometers were measured during both field work excursions.

An abandoned gravel pit, located 200 m to the south of the PARC site (Figure 3-1), was also visited in order to investigate the local geological framework. A series of photographs were taken in a north-south direction along a wall within the gravel pit (Figure 3-3 and Figure 3-4).
Figure 3.3: Gravel pit excavation photo. The white box at the north end indicates the location of Figure 3-4. Unit 1 is horizontally bedded coarse sand with gravel layers. Unit 2 is medium to coarse sand. Unit 3 is bedded gravel with sand and large cobbles that is dipping slightly to the south. Unit 5 is medium sand with some gravel and cobbles. Unit 5 is talus that has fallen off the excavation face. Upon closer examination (Figure 3-4), the heterogeneity is more complex than described above. Tire marks are visible in the bottom right part of section B.
Figure 3.4: Close-up of the area indicated in Figure 3-3. The interbedded nature of the coarse sand and gravel deposits is apparent. The pen circled in red is 17 cm long.

3.4.2 Borehole Logging

Natural-gamma and electrical conductivity logs were acquired for all piezometers at the site except for CDA2 (see Figure 3-2 for locations). The logs were collected using a Mount Sopris MGX-II portable digital logger, including a 305 m winch, a 2PGA-1000 gamma tool, and a Geonics 2PIA-1000 electromagnetic induction (conductivity) tool.
The natural-gamma logs were collected at 2.75 m/min in an upwards direction, with a sampling interval of 0.005 m. Resulting data were later filtered using a 21-point moving average filter. The conductivity logs were collected at 2.75 m/min in an upwards direction, with a sampling interval of 0.01 m. No processing was applied to the conductivity logs.

3.4.3 Ground Penetrating Radar

GPR data were collected using a Sensors & Software pulseEKKO 100 GPR unit at 100 MHz and 50 MHz. This GPR unit consists of separate transmitting and receiving antennas, a control box and a laptop. GPR lines were oriented such that they passed by the piezometers as closely as possible, but due to rows of raspberry bushes, they were generally restricted to the grass roadways between the test plots.

Both common-offset and common-midpoint data were collected as part of the GPR survey. A common-midpoint (CMP) sounding was conducted first along Line 1 at both 100 MHz and 50 MHz antenna frequencies (see Figure 3-2 for location). For the CMP sounding the initial antenna separation was 0.2 m, with an increase in separation of 0.2 m for each subsequent measurement. The rationale for these chosen separations are discussed in Appendix II.

Common offset surveys were conducted at both 100 MHz and 50 MHz on 7 lines (Figure 3-2). The lines formed a grid with line spacing varying between 20 to 190 m. Profile surveys at 100 MHz used a constant antenna offset of 1 m (based on the results of the CMP survey) with a station spacing of 0.2 m. Profile surveys at 50 MHz used a constant offset of 2 m with a station spacing of 0.4 m.
GPR data were processed using the software ReflexW (Sandmeier, 2005) as outlined below;

1. Time zero correction;
2. Dewow filtering (moving average windows of 30 ns and 15 ns for 50 MHz and 100 MHz data, respectively);
3. Gaining (energy decay);
4. Velocity Analysis (diffraction hyperbola fitting);
5. Constant Velocity Migration (at 0.100 m/ns);
6. Regaining and Depth Scale (using constant average velocity of 0.118 m/ns).

After applying a constant time-zero shift, a moving average dewow filter with a window length of 30 ns for the 50 MHz antenna data, and 15 ns for the 100 MHz antenna data was applied. This was followed by an energy decay gain. All of the profiles exhibited prominent near surface diffractions (apparently from boulders) that interfered with the stratigraphic image. Velocities determined from the diffraction hyperbolas ranged from 0.095 to 0.105 m/ns. These diffractions were therefore filtered by applying a constant velocity migration to the data using a velocity of 0.100 m/ns. There was a strong reflection from the water table at the PARC site, and the velocity determined from the known depth of the water table at the time of the surveys provided a slightly higher velocity of 0.118 m/ns. The depth scale on the profiles was calculated using this higher velocity.
3.5 Results

3.5.1 Borehole Logging

Below the surface organic soil layer, gamma counts ranging between 20 and 50 cps (Keys, 1997) (Figure 3-5 and Figure 3-6) indicate material of predominantly sands and gravels around the piezometers, which is consistent with the limited lithology data collected from the boreholes during drilling. Rudimentary grain size analyses indicate a predominance of sand and gravel in the piezometers at the PARC site. For a discussion of the uncertainty associated with the logging data, refer to Appendix III.

Figure 3-5 shows the conductivity and natural gamma logs for Piezometer 91-1, situated at the northwest corner of the site (see Figure 3-2). These logs are typical of the borehole logs collected at the site. All logs exhibit an increase in conductivity beneath the water table (note that the water table was measured using a water level meter in each piezometer at the time of logging). Below the water table there is a clear inverse correlation between conductivity and natural-gamma values, whereas above the water table there is no clear correlation. This increased conductivity in saturated, slightly coarser-grained material may be partly, or entirely, due to greater mobility of ions in larger and/or better connected pore spaces, rather than increased ion concentration. Determining the balance of these two roles (greater ion concentration, vs. greater mobility) would require targeted groundwater sampling within zones of higher and lower gamma counts to determine the respective groundwater chemistry; unfortunately, this is not possible with the existing piezometer placements. The relatively sharp rise in conductivity at the water table also suggests that the water table should produce a prominent radar reflection.
In the natural-gamma logs, fining upwards (and limited coarsening upwards) sequences can be seen in the sand and gravel (Figure 3-6). There are also sudden shifts in the gamma counts (either up or down) that, in the context of the depositional environment, suggest the presence of erosion surfaces or scour and fill features (Clague, pers. comm., 2006). The fining upwards sequences tend to be repeated on a scale of 3 to 10 m. These trends are evident across a lateral scale of up to 10 m, or more, as shown by the cross-section constructed from the natural-gamma logs from Piezometers 91-4, 91-5 and 91-7 (Figure 3-6). This cross-section suggests that not only
does heterogeneity exist in a vertical sense, but also in a lateral sense, on a scale of up to 10 m, the limit imposed by the piezometer separation. The lateral connectivity of these units across 10 m may produce permeable pathways that may act as conduits for groundwater movement and nitrate transport. It is likely that other continuous layers such as the ones observed in Figure 3-6 would be present at other nearby locations within the outwash deposits.

While coarsening upwards sequences do appear within the logs they are not as common as fining upwards sequences. The coarsening upwards sequences generally do not appear to have as significant a range in grain size as the fining upwards sequences, as is seen in Figure 3-6, and do not appear to be as laterally continuous as the fining upward sequences. To investigate whether this heterogeneity can be extended beyond the nested piezometers and across the site, GPR surveys were conducted.

Due to the high energy depositional environment in which these sands and gravels were deposited, and the proximity to their source, there was frequent erosion and deposition of subsequent layers. Variations in energy levels of the environment lead to the bimodal nature of the sands and gravels. This alternating between coarser and finer grained material is evident in the coarsening and fining upwards sequences observed in Figure 3-6. Erosional events (Clague, pers. comm., 2006) lead to the abrupt changes in the natural gamma count, as this reflects the abrupt changes in grain size that are observed at these unconformities.
3.5.2 Ground Penetrating Radar

Ground penetrating radar profiles (Figure 3-7 and Figure 3-8) exhibit a strong reflection between 235 and 320 ns which is interpreted to coincide with the water table. Based on measured water table elevations at the site at the time of the survey and approximate velocity-depth conversion, a velocity of 0.118 m/ns was determined. This value is slightly higher than the velocity of 0.100 m/ns determined by the diffraction hyperbola near the surface of the unsaturated aquifer material. Both of these values are typical of partially saturated sands. For a discussion of the error that may be associated with these velocities, refer to Appendix II. These velocities are
also similar to the average vertical unsaturated zone velocity of 0.12 m/ns determined by Irving and Knight (2003). The difference between the velocities may be explained by the seasonal variations in saturation levels above the water table. This survey was conducted in May at the end of the winter/spring rain, which would produce more highly saturated conditions than in August towards the end of the dry summer season when Irving and Knight completed their survey (Irving, personal communication, 2006).

Data for three of the lines are shown: Line 2 and Line 6 are approximately 165 m and 180 m long, respectively (Figure 3-7), and Line 3 is 385 m long (Figure 3-8). The top profiles show the 50 MHz data, and the bottom profiles show the 100 MHz data. On all three lines, there are reflections present other than the water table reflection, that are likely caused by slightly finer grained material having greater moisture retention.

Interpretation is based on reflection configurations described in Beres and Haeni (1991). All of the GPR profiles exhibit similar reflections and overall character. Many reflections exhibit an undulating to chaotic nature, with diffractions occurring predominantly close to the ground surface. The undulating character is dominant, suggesting that the formation is comprised of bedded sands and gravels typical of a high energy fluvial environment. In areas that appear more chaotic in nature, smaller scale (not fully resolved) sand and gravel cross-bedding and/or smaller lenses are likely present, with larger diffractions caused by boulders. Such features, including boulders, are observed in the gravel pit (Figures 3-3 and 3-4).
Depth (m) at v=0.118 m/ns

Figure 3-7: GPR lines collected along Line 2.a and b, and Line 6.c and d. Figure 3-7.a and 3-7.e are the 50 MHz profiles while Figure 3-7.b and 3-7.d are the 100 MHz profiles. Some prominent reflections noted for both frequencies are indicated with white arrows and the water table is indicated with black arrows. The piezometers that are adjacent to the GPR lines are indicated.
Figure 3.8: GPR lines collected along Line 3. Figure 3-8a is the 50 MHz profile and Figure 3-8b is the 100 MHz profile. Some prominent reflections noted for both frequencies are indicated with white arrows and the interpreted water table reflection is indicated with black arrows.
The undulating reflections are generally horizontal and parallel to sub-parallel. Most of these reflections are not laterally extensive, extending no more than 10 m to 15 m along the profile. Other reflection types are much longer, extending up to a few hundred meters. Two shaded areas are shown on Figure 3-7a, which represent the layering identified at the piezometer scale (as defined in Figure 3-6).

The slope of the water table shown in the GPR sections (Figures 3.7 and 3.8) is consistent with the gradient observed in the piezometers at the PARC site, as well as the water table gradient within the regional groundwater flow model. The groundwater at the PARC site flows northwest to southeast.

The photographs from the gravel pit (Figures 3-3 and 3-4) similarly show distinct layers of coarser and finer material. Thickness of the sand and gravel units ranges from 20 cm to approximately 3 m. Most of these layers are laterally continuous, from several meters up to greater than the extent of the photo. There are also smaller units that are only laterally continuous over a scale of 10's of centimetres to a few metres. Overall, the scale of the heterogeneity present vertically and laterally at this gravel pit is consistent with the results of the borehole geophysical logs and the GPR profiles. Where borehole geophysics resolves finer-scale vertical heterogeneity, primarily general layering is delineated using GPR.

3.6 Discussion

3.6.1 Geologic Context

The PARC site is located in the Abbotsford outwash within the Sumas Drift deposits. This outwash deposit was close to the ice margin and was a high energy glaciofluvial environment, resulting in highly variable braided streams, which changed course frequently (Clague, personal communication, 2006). The generally horizontal and sub-parallel depositional layers are often truncated by overlying material, which can be of varying grain size. At the gravel
pit, there is evidence of gravel bars, truncation of underlying units, as well as scour and fill features (Figures 3-3 and 3-4). The undulating to chaotic nature of the GPR reflections (Figures 3-7 and 3-8) indicates that this depositional variability extends across the PARC site and likely throughout the Abbotsford outwash deposit.

This depositional environment likely led to frequent erosion and deposition of subsequent layers and, ultimately, to the bimodal nature of the sand and gravel deposits, which is evident in the natural-gamma logs (Figure 3-3 and 3-4). The high energy of this depositional environment would also result in little to no fine-grained (e.g., clay / silt size) material remaining in the deposits. The natural-gamma logs support this idea, as the gamma count does not exceed 50 cps (the cut-off between sand and finer grained materials). As well, on the GPR profiles, with the exception of the water table, there are few strong reflectors, suggesting that there is little fine material at this site.

The geophysical results suggest that the subsurface at the PARC site is highly heterogeneous. However, the impact of this heterogeneity on groundwater flow and nitrate transport will depend largely on the scale of investigation and whether groundwater is flowing in the unsaturated or saturated zone, as discussed below.

3.6.2 Implications for Nitrate Transport

Groundwater nitrate concentrations have been monitored historically across the Abbotsford-Sumas aquifer, and are typically observed to decrease with increasing depth (McArthur and Allen, 2005). However, at several locations within the aquifer, such as the PARC site, long term monitoring in nested piezometers points to lower nitrate-nitrogen (NO₃-N) concentrations at shallow depths, and higher concentrations at intermediate and deeper depths (Figure 3-9).
At the southeast corner of the PARC site, NO$_3$-N concentrations at 91-7 are historically lower than those measured in the deeper piezometers (91-5 and 91-4), and the mid-depth piezometer (91-5) generally has slightly higher concentrations than the deepest one (91-4). Piezometer depths are 21.1 m (91-7), 30.65 m (91-5), and 46.0 m (91-4). The screen is located across the bottom 1 m of each piezometer, and the sample is drawn from mid-screen depth. Thus, nitrate concentrations are generally representative of the piezometers depth. The water table at this location, as measured in summer 2005, is approximately 21 m.

Land use at the site is primarily for raspberry crops, particularly in the vicinity of this piezometer nest, and the plants are fertilized primarily during the growing season (May - August). There is a good degree of correspondence of nitrate concentrations between all three piezometers, and previous work by McArthur and Allen (2005) demonstrated that concentrations across the entire aquifer vary on an annual basis, following the precipitation (and aquifer recharge) trends.

![Historic nitrate nitrogen concentrations at a piezometer nest located at the southeast corner of the PARC site. See Figure 3-2 for location of piezometers. Water table depth was approximately 21 m in summer 2005.](image)

Figure 3.9: Historic nitrate nitrogen concentrations at a piezometer nest located at the southeast corner of the PARC site. See Figure 3-2 for location of piezometers. Water table depth was approximately 21 m in summer 2005.
Application of nitrogen-based fertilizers and manure (also commonly used in the Abbotsford aquifer) will result in water and dissolved nitrate infiltrating the vadose zone, and moving downward toward the saturated zone. In the vadose zone it is assumed that there is predominantly vertical flow. With nitrate loading occurring at surface, one would anticipate that as the nitrate moves to greater depths, the concentrations should become diluted due to dispersion and mixing. However, at the PARC site, the shallow piezometers (91-3 and 91-7) have lower nitrate concentrations than the deeper ones, consistently over the historic period.

There are three likely explanations for the concentration inversion. First, and the most likely, is that the source of nitrate to the deeper piezometers may not be derived from the site immediately above. Rather, it is possible that high NO$_3$-N concentrations at depth are derived from adjacent land use activities, and that nitrate moves laterally beneath the study site in accordance with the local hydraulic gradient. Wassenaar et al. (2006) obtained ages ranging from 7.8 years to 14.7 years for the moderate and deep piezometers at this site, indicating larger ages than would be anticipated if the nitrate had been derived from immediately above. This is certainly a viable explanation for this and other similar sites. Second, the nature of the heterogeneity of the sediments may result in variations in measured concentrations, just by virtue of the grain size of the sediments at the completion depth of the piezometer. For example, if a coarse-grained unit is laterally continuous, as suggested by the results in Figure 3-5, then it may act as a permeable pathway for nitrate transport. However, to have such a consistent negative correlation between gamma and conductivity would seem to imply that this is more a result of a grain size change, with an associated greater ionic mobility within the larger pore spaces and not just the ionic concentration. Third, it is perhaps not coincidence that the lowest nitrate concentrations are measured in the two piezometers with completion depths at or slightly below the water table. The water table at this site fluctuates by approximately 2 m annually. Therefore,
it is reasonable to assume that there might be an increased potential for flushing nitrate within this depth range.

While it is possible to speculate on the reasons for the observed spatial distribution of nitrate at the site scale, the most reasonable approach would be to test the various scenarios by completing a numerical groundwater flow and transport model. However, in order to simulate nitrate transport in this, and similarly heterogeneous aquifers, consideration must be given to 1) the nature of the heterogeneity, and 2) the scale of the heterogeneity.

In the unsaturated zone, water and dissolved nitrate infiltrates slowly, and moves in a predominantly downward direction. Horizontal layering, present here as a result of the \(<5\) m scale fining upwards sequences, will result in enhanced vertical anisotropy in the hydraulic conductivity (i.e., lower \(K_v\)) due variable saturation of the coarse and fine grained materials. Vertical anisotropy will, thus, inhibit the downward movement of groundwater and dissolved nitrate. Therefore, in undertaking nitrate transport simulations within the vadose zone, it is important to take into account vertical anisotropy of the sediments.

In the saturated zone, groundwater flow occurs in a predominantly horizontal direction (note that there is a substantial horizontal gradient in most areas of this aquifer) and the effects of vertical anisotropy will be less pronounced compared to the effects of lateral continuity. For example, groundwater generally seeps more quickly though coarser grained horizons, with these more permeable horizons possibly acting as preferential pathways for nitrate transport. Therefore, in undertaking nitrate transport simulations within the saturated zone, the presence of laterally continuous geologic units should be taken into account if the scale of the heterogeneity is appropriate to the scale of the model.

Within the Abbotsford Aquifer and, particularly within the PARC site, two different scales of heterogeneity were identified. In the borehole logs, vertical heterogeneity is present as repeating fining upward sequences on the order of 3 to 10 m, and from the perspective of
modelling transport through the vadose zone, it is likely that the scale of the features can be readily incorporated into a vertical (one-dimensional) model. Modelling nitrate at the site (or larger) scale, however, is more problematic. Some of the fining upward sequences appear to be laterally continuous from one piezometer to the next over a distance of 10 m, and the GPR profiles indicate that bedding extends from a few meters to several hundred meters; well within the scale of the site. Therefore, at a site scale (less than a few hundred metres) the lateral continuity of coarser beds should reasonably be taken into account, as these may provide preferential pathways for nitrate transport.

At a larger scale, extending perhaps several hundred meters to kilometres beyond the site, it is unlikely that these sequences are laterally continuous, nor resolvable. Therefore, for larger scale transport modelling, it is unlikely that the heterogeneity could be represented. The most appropriate approach to modelling, therefore, is to consider the aquifer media homogenous (and perhaps anisotropic), and to represent the aquifer with bulk hydraulic properties, recognizing that model calibration to observed nitrate concentrations may be somewhat poor due to local heterogeneity effects. Characterization of smaller scale heterogeneity can improve definition of the statistical bounds in model calibration assessments over a range of scales, as was suggested by Flach et al. (2005).

3.7 Conclusions

Borehole logging data show evidence of fining upwards sequences, occasional coarsening upward sequences, and abrupt changes in grain sizes. Some of the fining upward sequences appear to be laterally continuous from one piezometer to the next, over a distance of 10 m. The coarsening upwards sequences do not appear to have the same degree of lateral continuity. The fining upwards sequences repeat vertically on a scale of about 3 m to 10 m. Consistency of the inverse correlation between the natural-gamma and conductivity below the water table indicates variation in the electrical conductivity of the groundwater could be largely
due to greater mobility of ions within larger pore spaces of coarser-grained material. The possible contribution of varying groundwater chemistry to the conductivity variation cannot be ruled out, however, without correlated chemical analysis of water samples.

GPR profiles indicate that there are bedded sands and gravels present at the PARC site, with the bedding extending from a few meters to several hundred meters. Photographs from a nearby gravel pit indicate that layering is present on a vertical scale of 10’s of centimeters thick, which is smaller than those indicated by the natural gamma logs. Layers of this scale are more commonly observed visually than in the natural gamma logs, which records contrasts in gamma activity, and not necessarily grain size.

The small scale nature of the heterogeneity observed at the PARC site will be an important control over vertical groundwater flow and nitrate transport in the unsaturated zone. At a site scale, the lateral continuity of layers may be significant as they may contribute to preferential pathways for nitrate migration. However, at a larger scale, it is unlikely that these sequences are laterally continuous, therefore, for transport modelling, it is unlikely that the heterogeneity could be reasonably represented.

3.8 Acknowledgments

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4 LOCAL SCALE MODELLING OF AQUIFER HETEROGENEITY

This chapter discusses the local nitrate transport modelling that was completed for a 20 km² area around the PARC site. The purpose was to establish how the heterogeneity that was observed within the aquifer can potentially affect the movement of groundwater and, ultimately, nitrate transport. A portion of this chapter has been submitted for publication in the Proceedings of the Joint Canadian Geotechnical Society-International Association of Hydrogeologists (Canadian National Chapter) Conference to be held in Vancouver, BC, October 2006. The paper is entitled “Investigating the Scale of Heterogeneity and Implications for Nitrate Transport, Abbotsford-Sumas Aquifer, BC and WA” (Chesnaux et al., 2006).

4.1 Introduction

The Abbotsford-Sumas aquifer, in the area around the PARC site, consists of heterogeneous aquifer media, characterized by predominantly horizontal to sub-horizontal layering. The scale of heterogeneity of the aquifer material was identified using GPR and borehole geophysics during a site scale investigation (Chapter 3). The geophysics indicated that the aquifer is comprised of inter-bedded sands and gravels, which is consistent with drillers’ logs and observations at a nearby gravel pit. The borehole logging indicates that there is layering in the aquifer media, which is characterized by predominantly fining upwards sequences, but also some coarsening upwards sequences and abrupt changes in grain size. The fining upwards sequences were identified on the scale of 3 to 10 m. The GPR sections show a hummocky reflection pattern, indicating that there are inter-bedded sand and gravels. Many of the reflectors have a small lateral extent of less than 15 m. However, a few reflectors appear to extend to at least the length of the GPR sections (several hundred metres).
The observed heterogeneity within the Abbotsford-Sumas aquifer has potentially an important role in groundwater movement. In the unsaturated zone, the observed layering and, in particular, the fining upward sequences can potentially result in a decreased $K_z$ relative to $K_{xy}$ over several metres. This increased vertical anisotropy would result in a decrease in the vertical groundwater flow velocity, affecting such factors as recharge rate and nitrate transport through the vadose zone. The significance of the range of grain size in the vadose zone, leading to anisotropy, is the subject of ongoing research by Chesnaux (pers. comm.).

In the saturated zone, groundwater flow is thought to be predominantly horizontal, at least at the local and regional scales and, therefore, an increase in the vertical anisotropy will have less of an impact on the groundwater movement than in the unsaturated zone where flow is predominantly vertical. The more permeable layers within the sequences may, however, act as preferential pathways for the groundwater flow and nitrate transport.

In order to assess the impact of the heterogeneity on groundwater flow and, ultimately, nitrate transport, numerical modelling was undertaken. A local groundwater flow model was created around the PARC site. This model was based on a regional groundwater flow model created previously by Scibek and Allen (2005). Both models are described in the following two sections. Several different approaches were attempted to representing the heterogeneity. Once the best-fit model was determined for the aquifer conditions, nitrate transport modelling was completed within the local model (Chapter 5).

4.2 Regional Groundwater Flow Model Development

A regional groundwater model for the Abbotsford-Sumas aquifer was developed by Scibek and Allen (2005). Both steady-state and transient models are available. The groundwater flow conceptual model was based on the hydrogeological framework of the Abbotsford-Sumas
aquifer. A brief description of the regional model is provided below as it forms the basis of the local scale model used for this thesis. Extents of the two models are shown in Figure 4.1.

The geological framework was developed from several thousand borehole lithology logs (drillers' logs) as well as several geological and hydrogeological reports (Clague et al. 1998; Cox and Kahle 1999; Halstead 1977; Hunter et al. 1998; Kahle 1991; Ricketts et al. 1993; Ricketts and Liebscher 1994). The complex depositional history of the Fraser lowland during the Wisconsin glaciation of the Pleistocene period resulted in complex sequences of deposits, predominantly diamictons and stratified drift, which infill a Tertiary bedrock valley. The Abbotsford-Sumas aquifer is mostly unconfined and is composed of glacial outwash deposits consisting of sands and gravels with some clay lenses, and is commonly referred to as the Sumas Drift. There is significant heterogeneity of the hydrostratigraphic units, which likely results in complex groundwater paths, particularly at a local scale (Scibek and Allen, 2005).

The aquifer is underlain by a glaciomarine stony clay deposit (Armstrong et al., 1965), which reaches ground surface to the west of the aquifer. This unit is thought to act as a regional aquitard. Laterally, the valley sediments are confined by the Tertiary bedrock surface, which outcrops as mountains on both sides of Sumas Valley, and as small outcrops south of the Nooksack River. The elevation of the Tertiary bedrock surface beneath the Pleistocene deposits of the lowland varies considerably, indicating pre-glacial erosional topography with large relief (Easterbrook 1969). A digital representation of the Tertiary bedrock topography was generated using deep borehole data, existing bedrock contour maps (Hamilton and Ricketts 1994), valley wall profiles, offshore bathymetric contours, and extrapolated cross-sections through the study area.
Figure 4.1: Location of the regional groundwater model (Scibek and Allen, 2005) and the local model (this study).

The traditional approach to identifying the hydrostratigraphy, based on the generation of discrete hydrostratigraphic layers from interpolated lithologies between boreholes (i.e., a layered paradigm), was not possible due to significant heterogeneity of the sediments. Instead, clusters of boreholes were examined, and the hydrostratigraphic zones mapped directly into Visual MODFLOW v. 3.1.84 (Waterloo Hydrogeologic Inc. 2004). This involved defining, on a model layer by model layer basis, property zones that correspond to similar hydraulic properties (i.e., hydraulic conductivity, \( K \) and storage coefficients, \( S_x \) and/or \( S_y \)). These were referred to by Scibek and Allen (2005) as K- and S-zones.
Each K-zone represented in the model (Figure 4.2) was then assigned a unique hydraulic conductivity (K) and specific storage (Ss) value, based on measured aquifer properties. There is extensive pumping test and specific capacity data for the US side of the model (Cox and Kahle, 1999), but sparse information (mostly constant discharge pumping tests on major production wells) on the Canadian side. Mean values were calculated from the composite dataset for each hydrostratigraphic unit mapped within the model. Data were related specifically to the aquifer material encountered at the well screen. The hydraulic properties were then assigned to the particular hydrostratigraphic units in MODFLOW, effectively providing estimates to areas with poor pump test data. This approach is reasonable given relatively small area of the aquifer, and the continuity of mapped hydrostratigraphic units on both sides of the border. K and Ss data are observed to have a heterogeneous distribution and strong zonation in some areas as illustrated in Figure 4.2.
Figure 4.2: Hydraulic conductivity zones for layer 1 within the regional model showing the extent of the local model within the regional model. White dashed line shows the location of the cross-sections shown in Figures 4-7 and 4-8.
During the calibration process there were zones that did not respond to reasonable adjustments of in K. The geology in these zones was re-interpreted from borehole lithologs, with attention given to model residuals, surficial geology, and individual borehole logs. Table 4.1 shows the final aquifer media hydraulic properties for each hydrostratigraphic unit in the model. The hydrostratigraphic zones in Table 4.1 were delineated based on the regional geology (Scibek and Allen, 2005).

Table 4.1: Final aquifer media hydraulic properties for each hydrostratigraphic zone in the regional model. See Figure 4.2 for zone distribution.

<table>
<thead>
<tr>
<th>Zone</th>
<th>$K_{xy}$ (m/d)</th>
<th>$K_{x}$ (m/d)</th>
<th>$S_{y}$ (1/m)</th>
<th>$S_{y}$ (-)</th>
<th>Effective Porosity (-)</th>
<th>Total Porosity (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>10</td>
<td>5</td>
<td>0.04</td>
<td>0.4</td>
<td>0.45</td>
<td>0.45</td>
</tr>
<tr>
<td>2</td>
<td>0.08</td>
<td>0.008</td>
<td>0.001</td>
<td>0.03</td>
<td>0.35</td>
<td>0.35</td>
</tr>
<tr>
<td>3</td>
<td>150</td>
<td>100</td>
<td>0.001</td>
<td>0.2</td>
<td>0.25</td>
<td>0.25</td>
</tr>
<tr>
<td>4</td>
<td>20</td>
<td>2</td>
<td>0.01</td>
<td>0.1</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>5</td>
<td>6</td>
<td>0.6</td>
<td>0.00255</td>
<td>0.05</td>
<td>0.35</td>
<td>0.35</td>
</tr>
<tr>
<td>6</td>
<td>3</td>
<td>0.3</td>
<td>0.001</td>
<td>0.08</td>
<td>0.35</td>
<td>0.35</td>
</tr>
<tr>
<td>7</td>
<td>5</td>
<td>2.5</td>
<td>0.001</td>
<td>0.05</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>8</td>
<td>50</td>
<td>35</td>
<td>0.01</td>
<td>0.15</td>
<td>0.25</td>
<td>0.25</td>
</tr>
<tr>
<td>9</td>
<td>60</td>
<td>40</td>
<td>0.0405</td>
<td>0.2</td>
<td>0.25</td>
<td>0.25</td>
</tr>
</tbody>
</table>

Model boundary conditions are based on both physical and hydrologic features. The lower model boundary corresponds to the bedrock surface, as the bedrock is considered relatively impermeable. The lateral extent of the regional groundwater model is shown in Figure 4.1. The model domain extends slightly beyond the Abbotsford-Sumas aquifer proper, particularly towards the northwest, in order to adequately represent the physical and hydrologic features that might serve as appropriate model boundary conditions. These include regional surface water divides to the west and north, and the bedrock outcrops to the east. Surface water divides are thought to approximate the regional groundwater divides as the aquifer is largely unconfined. The model domain extends beyond the Nooksak River to the south, in order to capture possible hydraulic influences from the region north of Bellingham, although the influence of this southern boundary
is small due to the high influence of the Nooksak River. Other model boundary conditions include the specified head and ditches, corresponding to the numerous streams that drain the aquifer and the major rivers that receive this drainage (i.e., the Nooksak and the Sumas Rivers). Head values for these features were determined using a combination of survey data and topographic information as described by Scibek and Allen (2005). Finally, recharge was modelled using the HELP software (US Environmental Protection Agency), and mapped spatially across the aquifer, taking into consideration the range of soil media types, shallow aquifer permeability and depth to water table (Scibek and Allen, 2006).

4.3 Local Groundwater Flow Model

The extent of the local model domain was determined by examining potential boundary conditions within the regional model, as shown in Figure 4.3. At the northern edge of the model, a groundwater high, corresponding to a local divide was used as a no-flow boundary. The eastern edge of the model was determined from the 48 m water table contour in the regional model, and was designated as a specified head boundary. The southern edge was a local groundwater divide that was also assigned as a no-flow boundary. The western side of the model was determined from water level contours within the regional model, and was set as a variable specified head boundary with elevations ranging from 55 to 62 m. During model calibration the specified head boundaries along the eastern and western boundaries of the local model area were lowered slightly to better reflect the local groundwater elevation conditions. The final elevations on the western side ranged from 48.5 to 61 masl, and 37 to 45 masl along the eastern side.
Figure 4.3: Water table elevation in regional groundwater model in the area of the local groundwater model.

The local model is located within the Abbotsford outwash deposits. The underlying Fort Langley Formation has a hydraulic conductivity three orders of magnitude less than the Abbotsford outwash deposit and, therefore, was used as a no-flow boundary at the base of the aquifer.

The calculated heads in the local model area were all elevated, suggesting that within the local model, some of the K values needed adjustment in order to recalibrate the model to local conditions. First, in the 3 zones with the highest K values (zones 3, 8 and 9) the K was decreased by one order of magnitude. Second, in the northern part of the model, the K in the low conductivity zone (Zone 2) was raised by one order of magnitude in all directions. Finally, in the
zones around Laxton and Judson lakes (Zone 1), K was decreased by 3 orders of magnitude to account for the elevated water levels in the wells immediately surrounding the lakes. These adjustments resulted in a better calibration between the observed and calculated heads.

Spatially-distributed mean annual recharge was used for this steady-state model. Since the recharge values for the model had been determined previously (Scibek and Allen, 2006), these were not adjusted within the model. However, to test the sensitivity of the local model to recharge, it was run under September base flow recharge conditions. Results indicate that reducing the recharge accordingly reduces the heads across the local model. Within the local model area, observed changes in groundwater elevation throughout the year, based on provincial observation well data, can be as great as 3 m (BC MOE, 2006). Thus, the aquifer is highly sensitive to recharge.

![Figure 4.4: Results of the model calibration for the local model. The NRMS is 9.635% and the correlation coefficient is 0.84](image)
The model was run under steady-state conditions, and calibrated using the observed historical static water levels in domestic wells, and taking into account the model water balance. The model calibration results are shown in Figure 4.4. When the Normalized RMS (root mean squared) is at 9.635 %, the water balance has a discrepancy of only -2.66 %. The NRMS of the model can be lowered, but when this occurs, the discrepancy in the water balance increases to an unacceptable level.

As validation of the local groundwater flow model, it can be compared with the regional groundwater model. Both models show similar flow properties are required on both the local and regional scale in order to represent the water levels observed in the aquifer. This model to model comparison increases the confidence in both flow models.

4.4 Representing Heterogeneity in the Model

When an aquifer is comprised of heterogeneous materials spanning more than one order of magnitude in K, such as is found in the Abbotsford-Sumas aquifer, there is always the question of how to best represent this within a numerical model. The geophysics conducted at the PARC site in Abbotsford indicated that heterogeneity is present on the scale of <1 to 5 m in the vertical direction and from a few to several hundred metres in the horizontal direction. Visual identification from the gravel pit photographs suggests that heterogeneity may also exist at scales finer than this; on the order of tens of centimeters. Several different approaches were taken to try and refine the hydrostratigraphy in the model such that heterogeneity might be better represented. Ultimately, an equivalent anisotropy approach was used.

4.4.1 Lithology Reassessment

Initial attempts to refine the model involved a re-examination of the borehole lithology logs present within the local model area. A total of 27 wells located in the area of the PARC site were re-examined (Appendix IV). It was thought that at the reduced scale of the model, relative
to the regional model, a greater degree of heterogeneity could be represented if a smaller and more manageable number of well logs were re-examined.

The gravel pit photographs (described in Chapter 3) indicated that there are many small scale variations in the geology. Although there were some indications of variations in the grain size within some of the logs, many simply indicated that the material is ‘sand and gravel’, with no information provided by the driller as to percentages of sand and gravel nor grain size (in the case of sand). The amount of fines present was similarly not recorded, due either to the fact that fines were not present or that the fines were washed from the samples. With the lack of detailed lithology information and the large distances between the borehole log locations, no further refinement of the hydrostratigraphy could be made. Thus, an alternative approach was sought.

4.4.2 Adding Homogenous Layers to the Model

Given the lack of success in refining the model through re-examination of the lithology data, a generic approach to representing this 5 m scale heterogeneity was attempted. This involved trying to represent the layering observed at the PARC site in the local model using the geophysics.

The initial approach to represent the model heterogeneity was to adopt a homogeneous layered approach and apply this to the shallow layers of the aquifer. The fining up sequence that was indicated in the geophysics appeared to repeat, on average, every 5 m. Therefore, the idea was to create a layered sequence, 5 m thick, and repeat it over the full model depth. To capture the fining upward character of the sequence, the 5 m interval was split into five 1 m layers, each assigned an isotropic K value, which decreased upwards through the layers in order to represent the observed decrease in grain size.

Two considerations had to be made in applying this approach. First, the model grid had to be refined to 1 m thickness. As there is a limit to the number of cells that can be created in
MODFLOW, this placed a restriction on the degree to which fine layering could be incorporated.

Second, representative properties needed to be assigned to each of the five layers. This second point proved to be quite challenging as only bulk hydraulic properties are available from pumping tests. The first approach to estimating the layered K distribution was using the gamma logs.

4.4.2.1 Grain Size Determination from Gamma Logs

As the grain size of a geologic material decreases, the observed gamma count generally increases (Keys, 1989). This relationship was examined in several of the piezometers at the PARC site. A crude grain size analysis was completed by Environment Canada on the “91” series piezometers at some of the depth intervals. An example from piezometer 91-1 is shown in Figure 4.5. The graphs from all of the piezometers with grain size analysis are included in Appendix V. In each of the intervals for which a grain size analysis was completed, the average grain size was calculated, and then plotted against the mean gamma count for the same depth interval (Figure 4.6). There appears to be no relationship, which is probably due largely to the large depth intervals over which the grain size analysis was completed and, therefore, the calculation of mean gamma. The depth intervals of the grain size samples were 4 ft (1.2 m), which in some cases is roughly equivalent to an entire fining upwards sequence. Due to the large sample interval, thin layers of finer or coarser material were not individually identified, but instead, blended with the material above and below it. If the sample interval was smaller, a better relationship may have been established from the gamma data.
Figure 4.5: Grain size distribution in piezometer 91-1.

Figure 4.6: Mean gamma count versus average grain size for piezometer 91-1.
4.4.2.2 Representing the Layered Sequence using the Equivalent K Approach

One method to represent layered heterogeneity is the equivalent K approach (Leonards, 1962). In this method, a series of homogeneous and isotropic layers can be represented by a single "equivalent" K value in each of the x/y and z directions. The equivalent K in the z direction is used if the flow is vertical, and in the x-y plane if flow is horizontal. Effectively, this leads to an anisotropic K distribution. Vertical anisotropy in the hydraulic conductivity values occurs when the $K_x$ and $K_y$ are equal but the $K_z$ has some other value, which is generally smaller than $K_x$. In this particular case, the calibrated regional and local models had been assigned anisotropic properties; therefore, the problem became one of working backwards to estimate the individual layer K values that might give rise to $K_z$.

A reasonable range of K values for sands and gravels were obtained from Freeze and Cherry (1979) that provided limits for the calculations of possible isotropic K values. The initial five K values were estimated within this range. Through an iterative process, these K values were modified until they provided equivalent K values to the anisotropic K values from the model.

Once the individual layer K values had been selected, the grid was refined to a 1 m thickness interval, and the properties assigned to each of the 5 repeating layers. The equivalent $K_{xy}$ of these layers was the same as $K_{xy}$ in the calibrated version of the model, and the equivalent $K_z$ of the layers was the same as the $K_z$ of the calibrated model. Figure 4.7 shows the model before the layering was attempted, and Figure 4.8 shows the layers applied to the model.
Figure 4.7: Cross section of local model without layering present.

Figure 4.8: Cross section of local model with layering present.
The approach proved to be impractical due to the fact that large, laterally extensive, but very thin layers were developed within the model. This is the direct result of the finite difference grid, which resulted in the same properties being assigned on a continuous basis across the entire model domain. Based on what is known about the lateral variability of the geology within the aquifer, this representation is not realistic. This type of depositional environment was not conducive to producing large flat sheets of material, and although some units can extend a few hundred metres, there is no indication that they are continuous over several kilometres.

Another difficulty encountered was accounting for the irregular grid at the model surface, due to topographic variations. In a finite difference grid, layers are added one atop the other. Unless a cell-by-cell adjustment is made to the layer top and bottom, the grid becomes very discontinuous where the layers shift to accommodate surface topographic variations. Manual adjustment of the cell properties proved to be very time consuming.

A third difficulty was the fact that grid refinement tends to be more pronounced in the upper layers of the model as this is where the more detailed stratigraphic data are available. However, from a groundwater flow perspective, many of the upper model layers are dry (the water table lies at considerable depth in some areas). Thus a lot of effort would be expended to modify cells, which effectively have no bearing on the flow and transport.

After considerable effort had been made to modify the model grid, it was determined that the approach would ultimately not lead to reasonable results and the approach was abandoned. It is worth mentioning, however, that if the model domain was smaller, the approach may have been feasible. Also, if a finite element code had been used, there may not have been the same level of difficulties making adjustments to the grid.
4.4.2.3 Modelling the Effect of Layering using a Small-Scale Model

Since using the layered approach at the scale of the local aquifer model was not possible, a 100 m by 100 m by 50 m test model was setup to determine what effect this layered approach would have on groundwater flow. The model consisted of 100 rows, 100 columns and 50 layers, resulting in cells that were 1 m x 1 m x 1 m (Figure 4.9). Three scenarios were established how layering affects particle tracking across the model. For all 3 scenarios, a generic 10 m gradient was applied from north to south, and particles were placed along the northern boundary of the model (Figure 4.9). MODFLOW (steady-state) and MODPATH were run for all 3 scenarios.

The first model scenario (Figure 4.9) was the most simplistic, and used vertical anisotropy to represent layered heterogeneity within the model. The $K_{xy}$ was set at 90 m/d, and the $K_z$ at 25 m/d (preliminary calibration values for the main sand and gravel zone in the local model). The results from Scenario 1 are shown in Figure 4.10. The particles travel from $y=100$ to $y=0$ along a uniform straight path, and take 3.5 days to travel the length of the model.

![Figure 4.9: Scenario 1 model. The model is 100m x 100m x 50m. The constant head boundary at y=100 is set at 50 m. At y=0, the constant head boundary is 40 m.](image-url)
Figure 4.10: Pathways (solid) in scenario 1 (homogeneous, vertical anisotropy) in the horizontal plane. Dotted lines are equipotential lines. The total travel time for the particles was 3.5 days.

The second scenario consisted of repeated 5 m fining upwards sets (Figure 4.11), similar to the layering attempted in the local Abbotsford-Sumas model. Within each set were 1 m thick isotropic layers that decreased in K upwards from one layer to the next. These sets were applied across the entire model. Using the equivalent K equations (Equations 4.1 and 4.2), the equivalent K_{xy} and K_z of these 5 m sets were set equal to K_{xy} and K_z in the first scenario. The K values for layers 1 through 5 were, respectively 6.5, 48.25, 90, 131.75, and 173.5 m/day.

\[
K_{xy} = \sum_{i=1}^{n} \left( \frac{K_i d_i}{d} \right)
\]  

[4.1]
Scenario 2 produced minimal change in the travel paths, as can be seen in Figure 4.12, but the travel time increased to 8 days. The layering had a significant effect on the travel times through the model. Particles moved slower through the low K zones, as compared with Scenario 1, resulting in the increased travel times.

\[
K_z = \frac{d}{\sum_{i=1}^{n} \left( \frac{d_i}{K_i} \right)}
\]  

Figure 4.11: Layering of hydraulic conductivity in Scenario 2. The K values for layers 1 through 5, respectively 6.5, 48.25, 90, 131.75, and 173.5 m/day were repeated throughout the model.
The third scenario was the most complex, and was undertaken to investigate the effect of discontinuous layering (Figure 4.13). The model was divided into 5 m by 5 m blocks, and the same layer as in scenario 2 was applied to the block in the upper left \((x=0, y=100)\) corner of the model. Moving either right or down, (increasing \(x\), decreasing \(y\)) the next block contained the same sequence, but all of the layers were shifted up by one relative to the block next to them. This systematic step-wise variation in the layering was continued across the model.

The greatest change in the travel path and travel times of the particles is observed this model scenario. Particles preferentially move through the higher \(K\), resulting in a distortion of the travel paths (Figure 4.14), and variations in the travel time of the particles. The particles take between 4.5 and 6.5 days to travel across the model. The lower average travel time reflects the presence of preferential pathways relative to scenario 2, where the particles are forced to travel through the full range of \(K\) values as they move from one side of the model to the other.
Figure 4.13: Variable layering of hydraulic conductivity in Scenario 3. The layered K values were 6.5, 48.25, 90, 131.75, and 173.5 m/day were repeated throughout the model.

Figure 4.14: Pathways (solid) for scenario 3 (varying layered blocks) in the horizontal plane. Dotted lines are equipotential lines. The total travel time for the particles ranged from 4.5 to 6.5 days.
The test models show that heterogeneity within an aquifer has a significant impact on the movement of groundwater. The layered heterogeneity resulted in travel times (or ages) more than twice the duration of those from the vertically anisotropic model. This would indicate that using vertical anisotropy to represent layered heterogeneity will result in an underestimation of the travel times of the particles. The heterogeneity also produced deviations in the travel paths of the particles. The discontinuous layering resulting in significant lateral and vertical movement of the particles, which were not observed in either the simple layered or vertically anisotropic model.

4.4.3 Representing Layering using Anisotropy

The lack of sufficiently detailed lithologic information, along with the limits of MODFLOW in regard to implementing the observed heterogeneity, meant that the heterogeneity could not be incorporated into the model as thin layers. Thus, an attempt was made to represent the observed heterogeneity using vertical anisotropy. Although it was shown in the previous section that this may not be an ideal approach as the ages may be underestimated. The calibrated regional and local model incorporated a certain degree of anisotropy within each of the K units. Since the aquifer heterogeneity is generally related to horizontal to sub-horizontal layering, this anisotropy approach likely represents the heterogeneity well within the groundwater flow simulations, as evidenced by the reasonable calibration and water balance results. More importantly, however, is the impact that this anisotropy will have on the transport of nitrate. Nitrate transport can be expected to be sensitive to heterogeneity (anisotropy), more so than groundwater flow itself, as advective transport and dispersion are dependent on subtle variations in the groundwater velocities. Thus, varying the anisotropy can be expected to influence transport times.

To determine the most representative set of hydraulic properties for the aquifer, a sensitivity analysis and a particle tracking analysis were completed. Ultimately, the best model
was selected based on a calibration to isotopic groundwater ages (Wassenaar et al., 2006) in combination with the overall head calibration and water balance results.

4.4.3.1 Sensitivity Analysis

In order to determine the sensitivity of the model to the conductivity anisotropy, a sensitivity analysis was completed. This involved varying the anisotropy present within each K-zone in the model and determining the effect on the model calibration and water balance. Sixteen different conductivity configurations were examined, as indicated in the Table 4.2.

For each scenario listed in Table 4.2, the MODFLOW 2000 and Zone Budget modules were run. This provided the calibration values and water balances presented in Table 4.3. Several of the calibration scenarios produced NRMS values that are lower than in the calibrated scenario. These all occur in scenarios where the hydraulic conductivities are becoming either increasingly isotropic or homogeneous, or both. However, in many of these scenarios the water budget is very poor. Based on the information provided in the lithology logs, it is unreasonable to expect that an increasingly isotropic and homogeneous model would provide a realistic representation of the geology. In addition, the regional model by Scibek and Allen (2005) did include some vertical anisotropy, which was required to regulate the heads in the model. The local model in this study also required vertical anisotropy to calibrate the model.

The effect of anisotropy was tested in several simulations. In some, the $K_z$ for all units was reduced, while in others, the $K_z$ was reduced only in the sand and gravel. In scenario $K_{sgd\downarrow}$, the $K_z$ value was decreased by one order of magnitude within the sand and gravel units only. This resulted in a very small increase in both the NRMS and the water budget. Once the $K_z$ value was decreased further, the NRMS and water budget both increased. As well, lowering the $K_z$ value in all of the units, not just the sand and gravel units, resulted in poor NRMS and water budget values.
Table 4.2: Conductivity scenarios for the sensitivity analysis.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Configuration</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kcal</td>
<td>Calibrated model (K_x, K_y, K_z lower – unique values for each zone)</td>
</tr>
<tr>
<td>Khomo</td>
<td>K_x=K_y=K_z in each conductivity zone with values same as those in calibrated model</td>
</tr>
<tr>
<td>Kallhomo</td>
<td>K_x, K_y, K_z in each zone same as the main sand/gravel zone (K_x=20 m/d, K_y=20 m/d, K_z=10 m/d)</td>
</tr>
<tr>
<td>Kallsame</td>
<td>K_x=K_y=K_z in all zones same as the K_x in the main sand/gravel zone (K_x=K_y=K_z=20 m/d)</td>
</tr>
<tr>
<td>K1down</td>
<td>K_x is reduced by 1 order of magnitude from Kcal in each zone</td>
</tr>
<tr>
<td>K2down</td>
<td>K_x is reduced by 2 orders of magnitude from Kcal in each zone</td>
</tr>
<tr>
<td>K3down</td>
<td>K_x is reduced by 3 orders of magnitude from Kcal in each zone</td>
</tr>
<tr>
<td>Khalf</td>
<td>K_x is reduced by 1/2 from Kcal in each zone</td>
</tr>
<tr>
<td>Khalfdiff</td>
<td>K_x is increased so that K_x-K_y is reduced by 1/2</td>
</tr>
<tr>
<td>Ksghomo</td>
<td>K_x=K_y=K_z in each sand/gravel conductivity zone</td>
</tr>
<tr>
<td>Ksgallhomo</td>
<td>K_x, K_y, K_z in each sand/gravel zone same as the main sand/gravel zone (K_x=20 m/d, K_y=20 m/d, K_z=10 m/d)</td>
</tr>
<tr>
<td>Ksgallsame</td>
<td>K_x=K_y=K_z in all sand/gravel zones same as the K_x in the main sand/gravel zone (K_x=K_y=K_z=20 m/d)</td>
</tr>
<tr>
<td>Ksg1down</td>
<td>K_x is lowered 1 order of magnitude from Kcal in each sand/gravel zone</td>
</tr>
<tr>
<td>Ksg2down</td>
<td>K_x is lowered 2 orders of magnitude from Kcal in each sand/gravel zone</td>
</tr>
<tr>
<td>Ksg3down</td>
<td>K_x is lowered 3 orders of magnitude from Kcal in each sand/gravel zone</td>
</tr>
<tr>
<td>Ksghalf</td>
<td>K_x is reduced by 1/2 from Kcal in each sand/gravel zone</td>
</tr>
<tr>
<td>Ksghalfdiff</td>
<td>K_x is increased to that K_x-K_y is reduced by 1/2 in sand/gravel zones</td>
</tr>
</tbody>
</table>

Table 4.3: Calibration results from the sensitivity analysis. The shaded simulations provided good NRMS and water balance results.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Normalized RMS (%)</th>
<th>Water Budget (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kcal</td>
<td>9.635</td>
<td>-2.66</td>
</tr>
<tr>
<td>Khomo</td>
<td>8.962</td>
<td>-95.16</td>
</tr>
<tr>
<td>Kallhomo</td>
<td>8.026</td>
<td>-164.44</td>
</tr>
<tr>
<td>Kallsame</td>
<td>8.011</td>
<td>0</td>
</tr>
<tr>
<td>K1down</td>
<td>10.509</td>
<td>-11.36</td>
</tr>
<tr>
<td>K2down</td>
<td>22.148</td>
<td>-3.58</td>
</tr>
<tr>
<td>K3down</td>
<td>22.305</td>
<td>-1.15</td>
</tr>
<tr>
<td>Khalf</td>
<td>9.893</td>
<td>-39.77</td>
</tr>
<tr>
<td>Khalfdiff</td>
<td>9.292</td>
<td>-3.8</td>
</tr>
<tr>
<td>Ksghomo</td>
<td>8.882</td>
<td>-0.28</td>
</tr>
<tr>
<td>Ksgallhomo</td>
<td>9.651</td>
<td>96.17</td>
</tr>
<tr>
<td>Ksgallsame</td>
<td>9.356</td>
<td>-100.49</td>
</tr>
<tr>
<td>Ksg1down</td>
<td>9.675</td>
<td>-2.67</td>
</tr>
<tr>
<td>Ksg2down</td>
<td>10.978</td>
<td>-32.88</td>
</tr>
<tr>
<td>Ksg3down</td>
<td>13.842</td>
<td>-20.22</td>
</tr>
<tr>
<td>Ksghalf</td>
<td>9.209</td>
<td>-88.6</td>
</tr>
<tr>
<td>Ksghalfdiff</td>
<td>9.215</td>
<td>-82.34</td>
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</tbody>
</table>
In the scenarios where $K_z$ was increased, thereby reducing the anisotropy, the calculated water levels within the model were lower than in the calibrated model. There was also a general reduction in the impact of pumping on the water table. When the model was made completely homogeneous and isotropic, scenario Kallsame, the groundwater flow was fairly uniform and only altered by the presence of constant head boundaries, such as the lakes, and the drains. When $K_z$ was increased only in the sand and gravel units, effects were seen across the entire model and not just where these units dominate.

In the scenarios where $K_z$ was decreased, resulting in increased vertical anisotropy, the overall water levels increased. Increasing the vertical anisotropy further resulted in a greater increase in water levels. This was particularly noticeable when $K_z$ was lowered within the lower K units. With the lower $K_z$, the effects of the pumping wells on the water table were increased. When $K_z$ was lowered only within the sand and gravel units, the water levels initially only rose in the sand and gravel dominated areas. However, as $K_z$ was further lowered, the water levels in the low K dominated areas also began to rise.

### 4.4.4 Particle Tracking

For each scenario used in the sensitivity analysis (see Table 4.2), particle tracking was completed to investigate the effect changing $K_z$ had on the travel length, time and velocity of particles placed within the model. The particles were placed around piezometers for which isotopic ages had been determined (Wassenaar et al., 2006), so that these ages could be compared with travel times determined from the numerical modelling. Eleven piezometers were identified within the model area that had isotopic ages. The locations of these piezometers are shown in Figure 4.15. A circle of 10 particles was placed at a 1 m radius around the piezometers at the average screen depth for each piezometer. There are 4 different sets of clustered piezometers with either 2 or 3 piezometers within a cluster.
The MODPATH module within Visual MODFLOW was run for each of the scenarios with backwards tracking particles to determine the travel path and age for all of the particles. The particle pathways for scenario Ksg1down are shown in Figure 4.16. From each of the scenarios, the travel time, lateral path distance, and resulting minimum velocities were determined with each set of particles.

Figure 4.15: Location of piezometers used for particle tracking.
In general, for the shallow wells (less than 5 m below the water table), changing the anisotropy had little affect on the travel time or length of the particle pathway. For particles placed at the deeper wells, there was a substantial change in the travel time and path length when the anisotropy was increased. The resulting particle velocities in these scenarios showed substantial decrease as the anisotropy increased. A decrease in anisotropy resulted in minimal change in the particle travel time. However, the pathway lengths generally increased, resulting in an increase in the velocity of the particles.

Figure 4.16: Particle pathways from scenario Ksgldown (Kz lowered by one order of magnitude in the sand and gravel units).
4.4.5 Comparison of Model Ages to Isotopic Ages

For each model scenario, the ages of the particles were plotted against isotopic ages determined by Wassenaar et al. (2006) using $^3$H/$^3$He dating techniques. Figure 4.17 shows this plot for the Ksg1down scenario. As can be seen in Figure 4.16, the particles tracking back from the piezometers at the western edge of the model terminate prematurely at the model boundary and, therefore, have not been included in Figure 4.17. These particles are shown on the graph, but are excluded from the calculation of the relationship between the model age and isotopic age.

![Graph of model age and isotopic age for scenario Ksg1down.](image)

Figure 4.17: Graph of model age and isotopic age for scenario Ksg1down.

In all of the 17 model scenarios that were run, there was a consistent underestimation of the age of the water particles, as compared with the isotopic ages. The model ages were generally between 60 and 80% of the isotopic ages. This discrepancy can be explained by the fact that the model is unable to take into account tortuosity of the water movement through the aquifer material. Figure 4.18 shows how tortuosity can effect travel times of groundwater. The model
representation of the aquifer material predicts a shorter travel path and time for movement through equivalent units.

![Diagram](image)

**Figure 4.18:** Tortuosity of groundwater results in a longer travel time. In A) the groundwater particle follows a tortuous path that would be found in aquifer material. In B) the model representation of the same material shows a shorter travel path and thus shows a shorter travel path and thus and shorter travel time.

The effect was also seen with the small test model in section 4.4.2.3. The numerical model will consistently predict shorter travel times (younger ages) that is observed in the aquifer material as it can not sufficiently account for the tortuosity of the groundwater movement.

By combining the results from the sensitivity analysis and the age determination, accepting that there in an inherent underestimation of the groundwater ages, it was possible to pick a most realistic hydraulic conductivity scenario for the aquifer. There were a number of scenarios presented in Table 4.3 that showed reasonable calibration values (shaded). It has already been stated that the increasingly homogeneous and isotropic scenarios do not provide a realistic representation of the geology present in the aquifer. When the quality of the age
comparison data was examined in addition to the model calibration for the various scenarios, it was determined that scenario Ksg1down provided the most realistic representation of both the geology and known water ages. This scenario had calibration values almost identical to the calibrated model and also a reasonable fit for the age data. The ages were determined by the model to be 77% of the isotopic age.
5 NO₃ TRANSPORT MODELLING

5.1 Model Input

Nitrate transport modelling was completed within the local groundwater flow model. Because MODFLOW is a saturated flow model, it was first necessary to estimate the concentration of NO₃ at the surface of the water table. To do this, several factors had to be considered.

First, as nitrate is considered to originate primarily from fertilizer applied to raspberries, an estimate of the nitrate loading at ground surface was needed. The nitrate is considered to be non-point source and distributed evenly across the field in which the berries are cultivated. To determine the spatial distribution of land surface nitrate sources, land use data were used (BC Ministry of Agriculture) (Figure 5.1). In addition to the application of fertilizers to berry fields, it is possible that there are other non-point sources (other crops) as well as point sources from manure stockpiles at poultry farms at some locations. Historically, these manure stockpiles were thought to contribute significantly to nitrate contamination in the aquifer. However, recent BMPs (beneficial management practices), along with the BC Agriculture Waste Control Regulation (BC Reg. 131/92, 1992) are expected to have reduced this source of contamination. Thus, only nitrate derived from the application of fertilizer to raspberry farms is considered in this study. The implications of this are discussed later.
Second, it was necessary to determine the concentration of nitrate arriving at the water table beneath each berry field. This is complicated for three main reasons:

1. The application timing and quantity applied have varied historically and, even today, there is considerable uncertainty with respect to both,

2. It is unclear how irrigation contributes to the mobilization of nitrate over the summer months and, certainly, there is little information on irrigation practices.
3. Nitrate transport through the vadose zone is dependent upon the aquifer media and the depth to the water table, both of which are variable across the aquifer. Within the model area, the water table depth varies from 0 to 35 m below ground surface.

Notwithstanding these uncertainties, models of the maximum concentration of nitrate arriving at the water table have been made based on average residual soil nitrate measured within the top 1 m of soil at the end of the growing season (September) over the past 4 years (Chesnaux et al., 2006). Vadose zone modelling was undertaken from October 1 to April 1 using mean monthly recharge, and representative soil and aquifer hydraulic properties (Chesnaux et al., 2006). Table 5.1 shows the maximum concentrations at the water table for different water table depths. The values represent a best estimate of current maximum concentrations, although historically, these concentrations are expected to have been higher (Allen, personal communication) due to over-fertilization.

<table>
<thead>
<tr>
<th>Water table depth (m)</th>
<th>NO₃ Concentration (mg N/L)</th>
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<tr>
<td>5</td>
<td>10.5</td>
</tr>
<tr>
<td>10</td>
<td>10.1</td>
</tr>
<tr>
<td>15</td>
<td>10.4</td>
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<tr>
<td>20</td>
<td>11.9</td>
</tr>
<tr>
<td>25</td>
<td>18.6</td>
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</table>

Because a steady state flow model is used, and mean annual recharge is applied to that model, these values were assigned as constant concentration boundary conditions within the model to represent annual loading. It was assumed that the groundwater and NO₃ movement in the unsaturated zone is vertical. The location of these constant concentration boundary conditions within the model were determined using a combination of the modelled depth to the water table, the model layer in which the water table occurs (layers 1 through 5), and the distribution of
raspberry fields in the model area. For each model layer, a distribution of nitrate at the water table was determined (Figures 5.2 – 5.6).

Figure 5.2: NO₃ constant concentration distribution in model layer 1.
Figure 5.5: NO$_3$ constant concentration distribution in model layer 4.

Figure 5.6: NO$_3$ constant concentration distribution in model layer 5.
In order to complete the setup of the model, representative transport parameters were assigned (Table 5.2). These include longitudinal, horizontal and transverse dispersivity, the diffusion coefficient, and effective porosity. There is a scale affect associated with the dispersivity values (Anderson and Woessner, 2002); however, the values used were appropriate for a model of this size. Due to the highly permeable nature of the aquifer, NO₃ transport is anticipated to be predominantly controlled by advection and, therefore, diffusion is expected to be insignificant.

Table 5.2: NO₃ transport model input parameters.

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<tr>
<th>Parameter Name</th>
<th>Input value</th>
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<td>Longitudinal Dispersivity¹</td>
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<td>Vertical Dispersivity¹</td>
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<tr>
<td>Effective Porosity³</td>
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</table>

¹Gelhar et al., 1992; ²Freeze and Cherry, 1979; ³Scibek and Allen, 2005

Piezometers that have NO₃ concentration data for the years 2002-2004 were used as calibration data for the transport model. These piezometers were situated on the Canadian side of the aquifer and monitored by Environment Canada. Data provided by Mitchell (2005) were used for calibration locations for the US side, as sampling was completed during the same years. The locations of the concentration observation piezometers are indicated in Figure 5.7. Concentration values are provided in Table 5.3.
Figure 5.7: Location of concentration observation piezometers.
Table 5.3: \( \text{NO}_3 \) calibration concentration values.

<table>
<thead>
<tr>
<th>Well ID</th>
<th>( \text{NO}_3 ) Conc. (mg N/L)</th>
<th>Well ID</th>
<th>( \text{NO}_3 ) Conc. (mg N/L)</th>
<th>Well ID</th>
<th>( \text{NO}_3 ) Conc. (mg N/L)</th>
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</table>

5.2 Model Scenarios

Three scenarios were modelled using the input parameters presented above. In Scenario 1, the initial \( \text{NO}_3 \) concentrations were set to 0 mg N/L across the entire model domain. This scenario was designed to investigate how the nitrate moves through the aquifer.

Scenario 2 was designed to investigate whether the nitrate concentrations suggested to be reaching the water table since the change in BMPs in the mid 1990s (Sutherland, personal communication, 2006) would produce the \( \text{NO}_3 \) concentrations observed currently within the aquifer. The 1992-1994 \( \text{NO}_3 \) spatial distribution map (see Chapter 2) was used for the initial
concentration distribution within the model. Figure 5.8 shows this distribution within the model area. The US part of the distribution was obtained by extrapolating the observed nitrate concentrations beyond the edge of the observed data. The data in Chapter 2 indicates that the concentrations of \( \text{NO}_3 \) drop off at depths greater than 50 m; therefore, the initial concentrations were only applied in approximately the upper 50 m (layers 1 through 7) of the model. Ideally, a separate concentration map would have been produced for each model layer, but because only a limited number of piezometers were available for each depth range, this was not possible. Nevertheless, because the piezometers are generally shallow, an integrated map was assumed to be a reasonable approximation.

In scenario 3, the current (2002-2004) \( \text{NO}_3 \) spatial distribution map (see Chapter 2) was used as the initial concentration data (Figure 5.9). This scenario was intended to examine what the \( \text{NO}_3 \) concentrations would be in 10 years, assuming that the currently modelled \( \text{NO}_3 \) concentrations reaching the water table were to continue into the future. Again the initial concentrations were only applied to a depth of 50 m.

All three scenarios were run using the MT3DMS module within Visual MODFLOW (Waterloo Hydrogeologic Inc., 2004) with the solution method parameters indicated in Table 5.4. All models were run for a total of 3650 days (10 years).
Figure 5.8: Initial 1992-1994 NO$_3$ concentrations. Black squares indicate the concentration observation locations. Line A-A' is dotted and shows the location of the cross-sections in Figures 5.10, 5.14, and 5.22. Line B-B' is dashed and shows the location of the cross-section in Figure 5.11.
Figure 5.9: Initial 2002-2004 NO₃ concentrations. Black squares indicate the concentration observation locations. Line A-A' is dotted and shows the location of the cross-sections in Figures 5.10, 5.14, and 5.22. Line B-B' is dashed and shows the location of the cross-section in Figure 5.11.
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<th>Method of Characterization</th>
<th>Fourth-order Runge-Kutta for sink and sources; Euler, elsewhere</th>
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</tbody>
</table>

101
The Peclet number, Pe, was calculated for the model using Equation 5.1. Using the grid size spacing (50 m) and the longitudinal dispersivity coefficient (100 m), the Peclet number was determined to be 0.5. With the Peclet number \( <1 \), this minimizes the errors associated with spatial discretization.

\[
P_e = \frac{\Delta l}{\alpha}
\]  \[5.1\]

where \( \Delta l \) is nodal spacing and \( \alpha \) is the characteristic dispersivity.

5.3 Results

5.3.1 Scenario 1: No Initial Concentrations

Results for Scenario 1 were used to show both the spatial distribution of \( \text{NO}_3 \) resulting from application at the water table and how the concentrations change with time. Figure 5.10 shows a cross section in a north-south direction through the aquifer showing both the initial concentrations and the concentrations after 10 years. When the \( \text{NO}_3 \) first enters the groundwater, it begins to move downwards through the aquifer. As time progresses, nitrate plumes spread along the general groundwater flow direction from north-west to south-east.

Concentrations within the wells increase gradually from an initial concentration of 0 mg N/L. After 10 years, the calculated concentrations are well below the observed \( \text{NO}_3 \) concentrations in the aquifer. The concentrations in shallow piezometers (\(< 20 \) m depth) increased to a maximum of approximately 10 mg N/L within the first 3 to 5 years of the simulation (Figure 5.11). Piezometers located deeper (20 – 45 m depth) took longer to attain an equilibrium concentration, with a few not reaching equilibrium after 10 years (Figure 5.12).
Figure 5.10: North-south cross-section showing nitrate distributions a) initially (0 years) and b) after 10 years in Scenario 1. Vertical exaggeration is 5x.
Figure 5.11: Scenario 1 NO₃ breakthrough graph for three piezometers located less than 20 m deep.
Figure 5.12: Scenario 1 NO$_3$ breakthrough graph for four piezometers located between 20 and 45 m depth.

An example of the effects of heterogeneity on NO$_3$ transport is shown in Figure 5.13 from Scenario 1. This example of large scale heterogeneity shows a low K unit in which there is minimal movement of the NO$_3$ released from the constant concentration boundary at the water table. Once the NO$_3$ moves into the higher K unit below the low K unit, there is significant lateral movement. This type of effect was observed in all of the low K units within the model.
Figure 5.13: Movement of NO$_3$ around a low K zone (white shaded), shown at 10 years. Vertical exaggeration is 10x.
5.3.2 Scenario 2: 1992-1994 Initial NO₃ Concentrations

Scenario 2 was designed to simulate the effect of the nitrate loading from residual nitrate concentrations measured in September on current nitrate concentrations. Implicit in this Scenario is the assumption that the only source of nitrate is that of residual nitrate present in the soil in berry fields at the end of the growing season. If the residual nitrate concentrations are representative of conditions arising from the new BMPs initiated in the early 1990s, then after 10 years, the observed concentrations should match the observed concentrations for 2002-2004. Figure 5.14 shows a cross-section through the same north-south column as in Scenario 1 showing the changes between 1992-1994 and 2002-2004.

Based on the NO₃ input concentrations suggested by the BMPs, it would be expected that the observed concentrations within the aquifer would fall as the lower concentration NO₃ flushes through the groundwater. The decreased final model concentrations agree with this. However, as shown previously in Chapter 2, the 2002-2004 observed NO₃ concentrations are generally higher than those of 1992-1994, suggesting that this is not actually the case. The breakthrough curve in Figure 5.15 for the shallow piezometers shows that regardless of the initial concentrations, the calculated concentrations still reach equilibrium within the first 5 years, as suggested by Scenario 1. Figure 5.16 shows the breakthrough curve for the deeper piezometers. These piezometers take longer than the shallow piezometers to reach equilibrium. In both the shallow and deep piezometers, there are some piezometers that show an increase in NO₃ concentration and others that show a decrease. Piezometers with lower initial concentrations, but a closer proximity to the NO₃ sources will show an increase, while those with high initial concentrations show a decrease in the calculated concentrations with time.
Figure 5.14: North-south cross-section showing nitrate distributions a) initially (1992-1994) and b) after 10 years in Scenario 2. Vertical exaggeration is 5x.
Figure 5.15: Scenario 2 NO₃ breakthrough graph for three piezometers located less than 20 m deep.
Figure 5.16: Scenario 2 NO$_3$ breakthrough graph for four piezometers located between 20 and 45 m depth.

The calibration graph shown in Figure 5.17 indicates that in piezometers with observed elevated NO$_3$ concentrations, the concentrations predicted by the model are up to 75% less than the observed concentrations after 10 years.
Figure 5.17: Calibration curve for Scenario 2 at 10 years. The NRMS is 22.827 % and the correlation coefficient is 0.334.

In Chapter 4 it was shown that the model ages were 77 % (Figure 4.17) of the isotopic ages determined for the groundwater. This would indicate that the groundwater is moving faster in the model than in the aquifer. Therefore, an examination of the model at 77 % of the 10 year model run length (7.7 years) may be more indicative of the actual NO₃ concentrations after 10 years in the aquifer. Figure 5.18 shows the calibration graph Scenario 2 at 7.7 years. At higher observed concentrations, the calculated NO₃ concentrations are considerably lower. They are almost identical to those at 10 years (Figure 5.17), as the concentrations appear to only change minimally during the last 5 years of the model run (Figures 5.15 and 5.16).
Layer #2: Conc001
Layer #3: Conc001
Layer #4: Conc001
Layer #5: Conc001
Layer #6: Conc001
Layer #7: Conc001
Layer #8: Conc001
--- 95% confidence interval
95% Interval

Figure 5.18: Calibration curve for Scenario 2 at 7.7 years. The NRMS is 22.434 % and the correlation coefficient is 0.371.

Most of the piezometers with observed elevated NO₃ concentrations are found in layer 4; however, there seems to be no relationship to the depth of the piezometer, or to the land use above the piezometer that would indicate why these concentrations are elevated. In all of these cases, the model predicts that these piezometers would again have NO₃ concentrations up to 75% less than the observed concentrations.

There are a few possible explanations for the low concentrations predicted by the model compared with the observed NO₃ concentrations. First, the constant NO₃ concentration boundary at the surface of the aquifer is based on assumed current BMPs, which have only been in effect since the mid-90s. Since it was determined in Chapter 4 that the groundwater reaching some of these piezometers is greater than 20 years old, the concentrations observed may indeed reflect older farming practices.
Second, the concentrations used for the constant concentration boundary are based on the concept that the only source of NO$_3$ to the aquifer is from the residual NO$_3$ soil concentrations at the end of the growing season (September). The low NO$_3$ concentrations calculated by the model would suggest that this source alone is not enough to produce the observed NO$_3$ concentrations. This would suggest that larger concentrations of NO$_3$ exist. This may involve infiltration of NO$_3$ at other times, such as during summer irrigation. The timing of fertilizer application with respect to rain events and irrigation may also effect the concentration of NO$_3$ reaching the aquifer. There may also be other point and non-point sources of NO$_3$ related to agricultural practices in the region, such as crop farming other than raspberries, along with poultry and livestock operations.

Third, the initial condition for this Scenario used a map of the concentrations from a depth-integrated dataset. That is, all concentration data, regardless of depth, were used to produce the map. Generally, the aquifer observed higher concentrations of NO$_3$ at shallower depths. Therefore, applying the vertically-averaged map to all model layers will result in possibly an over-estimation of initial nitrate concentration at depth. However, this would result in the model potentially over-estimating the nitrate concentrations, which appears not to be the case. Therefore, use of the depth-integrated map to represent initial conditions in the aquifer is reasonable.

It is likely that the poor calibration results are the result of a combination of historically high nitrate concentrations, and a model loading scenario for NO$_3$ that is not representative of actual conditions.

5.3.2.1 Dispersivity Sensitivity Analysis

A sensitivity analysis was completed within Scenario 2 to investigate the effect of varying dispersivity on transport. Results are compared to Scenario 2 (as described in section
5.3.2). In Scenario 2A, the dispersivity values were increased by one order of magnitude, and in Scenario 2B, they were decreased by one order of magnitude (Table 5.5).

Table 5.5: Values used in dispersivity sensitivity analysis.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Longitudinal Dispersivity</th>
<th>Horizontal Dispersivity</th>
<th>Vertical Dispersivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>100</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>2A</td>
<td>1000</td>
<td>100</td>
<td>10</td>
</tr>
<tr>
<td>2B</td>
<td>10</td>
<td>1</td>
<td>0.1</td>
</tr>
</tbody>
</table>

The results (Figure 5.19) show that when the dispersivity is increased by one order of magnitude, as in Scenario 2A, the nitrate travels further through the aquifer and becomes more widespread. When the dispersivity is decreased, as in Scenario 2B, the nitrate movement is restricted with minimal spreading observed.

The calibration results for Scenario 2A are shown in Figure 5.20 and indicate a minimal change in the calibration of the model. The majority of the calculated data are again clustered around 10 mg/L. In Scenario 2B (Figure 5.21) the data shows a considerably wider range of calculated concentrations. However there is little change in the actual calibration of the model. These results suggest that dispersivity is not a significant controlling factor in model calibration at this site.

5.3.3 Scenario 3: 2002-2004 Initial NO₃ Concentrations

Scenario 3 predicts what would happen to the initial 2002-2004 NO₃ concentrations, 10 years hence, assuming that the suggested NO₃ loading concentrations based on the current BMPs were to continue. Figure 5.22 shows the present day and 10 year NO₃ concentration distributions along the same north-south cross-section as in Scenarios 1 and 2.
Figure 5.19: Results from dispersivity sensitivity analysis. Figure A) shows the results from Scenario 2A where the dispersivity was increased by one order of magnitude. Figure B) shows the results from Scenario 2B where the dispersivity was decreased by one order of magnitude.
Figure 5.20: Calibration curve for Scenario 2A at 10 years. The NRMS is 23.911 % and the correlation coefficient is 0.120.

Figure 5.21: Calibration curve for Scenario 2B at 10 years. The NRMS is 21.826 % and the correlation coefficient is 0.372.
Figure 5.22: North-south cross-section showing nitrate distributions at a) the model start (2002-2004) and at b) 10 years in Scenario 3. Vertical exaggeration is 5x.
The results are similar to those in Scenario 2, with the NO₃ concentrations decreasing substantially in the first few years. Figure 5.23 shows the breakthrough curve for shallow piezometers and Figure 5.24 shows the same for deep piezometers. Again, the NO₃ concentrations reach equilibrium more quickly in the shallow piezometers than in the deeper ones. Some of the deeper piezometers do not reach equilibrium in the 10 years of the model.

Figure 5.23: Scenario 3 NO₃ breakthrough graph for three piezometers located less than 20 m deep.
In Scenarios 2 (Figure 5.14) and 3 (Figure 5.22) the NO₃ distributions after 10 years are very similar, although they have different initial NO₃ concentration distributions. Both model scenarios would, therefore, be expected to reach the same equilibrium NO₃ concentration distribution. Generally, however, the results from Scenario 2 indicate that there is an underestimation of the NO₃ reaching the aquifer; therefore, it is unlikely that Scenario 3 provides a realistic prediction of what will happen over the next 10 years.
5.4 Discussion

There are several factors that contribute to uncertainty in the transport simulations. First, there were no tracer tests completed in the aquifer, which could be used to verify the results, at least at a small scale. Second, there were only 11 isotopic ages available to compare with the model ages, and these are somewhat poorly distributed within the model. Also, there are inherent errors associated with dating techniques (presence of a mixed water) may contribute to error. As well, some of the ages for the deeper piezometers are greater than 10 years (Wassenaar et al., 2006); it may take longer than the 10 year model run to see significant changes in the deeper piezometers.

Perhaps the most significant source of uncertainty is the limited understanding of the nitrate loading at the surface of the aquifer. Clearly, the uncertainties surrounding the NO₃ loading at the water table will limit attempts to model NO₃ transport. Most of this uncertainty derives from lack of data on the timing and amount of fertilizer application and, specifically, the degree of compliance with current BMPs. In addition, however, the role of irrigation in mobilizing nitrate early in the growing season is uncertain. If NO₃ is mobilized as a result of irrigation, then the fact that residual NO₃ concentrations in the soil are low in September should not be used to make assumptions that current BMPs are, in fact, beneficial. Also, an assumption was made in the modelling that the only source of NO₃ is fertilizer application to berry fields. Other non-point sources (other crops) and point sources (from manure stockpiles) were not considered. These may provide additional widespread sources of NO₃ to the aquifer. Chesnaux (pers. comm.) is examining the nitrate loading in the Abbotsford-Sumas aquifer as part of ongoing research.
6 CONCLUSIONS AND RECOMMENDATIONS

6.1 Nitrate Distribution

In the Canadian portion of the Abbotsford-Sumas aquifer, groundwater nitrate concentrations are generally elevated above the normal background concentration of 3 mg N/L. Many areas have concentrations above the Maximum Allowable Concentration (MAC) of 10 mg N/L; in particular, the area to the southeast of the Abbotsford International Airport. The average concentration from 1998 to 2004 was 14±9 mg N/L. These elevated concentrations are reportedly caused by the use of fertilizers on the many farms in the area, particularly raspberries.

At groundwater depths less than 50 m there is little variation in the average nitrate concentration at each depth interval, which ranges from 12 to 16 mg N/L. Below 50 m, nitrate concentrations drop to below the MAC in most piezometers. There is no systematic increase or decrease in the nitrate concentrations with depth, which may be explained by the aquifer heterogeneity. Water that infiltrates from source areas with high nitrate loading may have followed preferential flow paths, which have produced elevated results at some depths and not at others. Although aquifer heterogeneity is an important factor, spatial variations in the NO₃ sources and timing of fertilizer application may also contribute to the observed variations.

Trends in nitrate concentration for the entire historic database suggest that since 1970, concentrations have increased. For the period 2002-2004, the trend is negative, although this appears to follow the seasonal variations in water levels. Changes in the spatial distributions of nitrate were determined by mapping available nitrate concentrations for different time periods. Over the past 10 years, changes have been both positive and negative, with some evidence of higher concentrations to the southeast of the Abbotsford Airport. Thus, despite efforts on the part
of the agriculture industry to reduce nitrate loading since 1990, there has not been a significant widespread reduction in nitrate concentrations. Hii et al (in draft) suggest that there may be a reduction in the last few years in shallow piezometers, however, this may be related to a lower water table observed in the aquifer during this time.

A recommendation following this study is that any available groundwater chemistry data from the U.S. portion of the Abbotsford-Sumas aquifer should be added to the chemistry database, and the appropriate graphs and maps updated, if possible, in order to have a more complete analysis of nitrate concentrations across the entire aquifer. Similarly, many of the wells for which chemistry data exist are missing well location and well completion information. This information should be compiled and the database updated.

6.2 Geophysical Surveys

Borehole geophysical logging and GPR proved to be useful for investigating the scale of heterogeneity at the well and field-site scales. Borehole logging data show evidence of layering defined by fining upwards sequences, occasional coarsening upwards sequences, and abrupt changes in grain sizes. Some of the fining upward sequences appear to be laterally continuous from one piezometer to the next, over a distance of 10 m, and repeat vertically on a scale of about 3 m to 10 m. The coarsening upwards sequences do not appear to have the same degree of lateral continuity. Consistency in the inverse correlation between the natural-gamma and conductivity indicates variation in the electrical conductivity of the groundwater, probably due largely to greater mobility of ions within larger pore spaces of coarser-grained material. The contribution of varying groundwater chemistry to the conductivity variation cannot be ruled out without correlated chemical analysis of water samples.

GPR profiles indicate that there are bedded sands and gravels present at the PARC site, with the bedding extending from a few meters to several hundred meters. This layering correlates
with the photographs from a nearby gravel pit, which also indicate that there is layering present on a vertical scale of 10's of centimetres thick. This vertical layering thickness is thinner than indicated by the natural gamma logs. Layers of this scale are more commonly observed visually than in the natural gamma logs, which records contrasts in gamma activity, and not necessarily grain size.

The small scale nature of the heterogeneity observed at the PARC site is an important control over vertical groundwater flow and nitrate transport in the unsaturated zone. At a site scale, the lateral continuity of layers may be significant as they may contribute to preferential pathways for nitrate migration. However, at a larger scale, it is unlikely that these sequences are laterally continuous, therefore for transport modelling, it is unlikely that the heterogeneity could be reasonably represented.

6.3 Modelling of Aquifer Heterogeneity

A local scale groundwater flow model was developed for the region surrounding the PARC site, using hydraulic parameters and internal boundary conditions from a regional scale model developed previously. Several approaches were taken to represent the aquifer heterogeneity within this local model. Refinement of the lithology within the model domain was unsuccessful due to the general lithologic descriptions provided in drillers’ logs. An attempt was also made to reconcile borehole grain size information with the gamma logs; however, this too was unsuccessful due to poor spatial resolution of the grain size data.

As indicated from borehole geophysical logs and gravel pit photographs, layering was characterized by repeated fining upward sequences on the scale of approximately 5 m. An attempt was made to represent these sequences conceptually by splitting each sequence into five 1 m layers with varying K values, and repeating these over several tens of metres within the model domain. A small test model indicated that this type of layering increases the travel time (or age)
of particles placed within the model, but has little effect on the total travel path. This technique was not transferable to the local scale model, due to the fact that the fining upward sequences are observed to be laterally continuous over 10 metres (the distance between piezometers). Thus, at the scale of the local model (20 km²), adding these continuous layers yielded an unrealistic representation of heterogeneity for this type of geological environment (glacial outwash).

The use of vertical anisotropy for hydraulic conductivity, \( K_z \), proved to be the most realistic approach for representing heterogeneity in the model. A sensitivity analysis was undertaken to determine the effect of varying \( K_z \) on model calibration and water balance. This was followed by particle tracking simulations to determine travel times, and comparing these to the groundwater ages derived from isotopic age dating of groundwater. Results suggest that generally one order of magnitude lower \( K_z \) than \( K_x \) provides the best relative calibration. However, the groundwater ages from the model were consistently 60-80 % of the ages determined using isotopic methods, likely due to the fact that flow path tortuosity is not well represented in the model. Another possible reason for the discrepancy between particle travel times and groundwater ages relates to the use of anisotropy to represent the heterogeneous aquifer material. In the small test model, the layered isotropic model resulted in longer travel times than the homogenous anisotropic model. Thus, within the local model, travel times may be similarly underestimated.

### 6.4 NO\(_3\) Transport Modelling

Nitrate transport modelling was undertaken in the local scale model. Constant concentration polygons were assigned in the model according to the distribution of raspberry fields. The maximum concentrations of nitrate arriving at the water table were determined from vadose zone transport of residual nitrate in the soil zone to different water table depths (Chesnaux, pers. comm.). An assumption was made that the only source of nitrate to the aquifer
is that of residual nitrate present in the soil at the end of the growing season. Using the nitrate loading data, three scenarios were modelled.

Scenario 1 (initial nitrate concentration equal zero) indicated that the NO$_3$ first moves down into the aquifer and then follows the general groundwater flow direction of north-west to south-east. It takes several years for the concentrations within the aquifer to reach equilibrium concentrations. Deeper wells and those further away from the NO$_3$ sources take longer to reach equilibrium.

Scenario 2 (initial nitrate concentration equal 1992-1994 nitrate distribution) showed that the simulated NO$_3$ concentrations are up to 75% lower than the observed concentrations in the aquifer after 10 years (the 2002-2004 data were used for calibration). There are four possible explanations for this. First, the modelled source NO$_3$ concentrations (based on residual NO$_3$ measured in the soil zone) may not be sufficiently elevated. Second, NO$_3$ loading may occur at times other than at the end of the growing season. This could be either during spring rainfall events (prior to plant growth) or during the summer as a result of irrigation. A third possibility is that there are other sources of NO$_3$, such as pastures, livestock farms and poultry farms contributing to the nitrate loading. Finally, current concentrations of nitrate may reflect historically high nitrate loading as some groundwaters have ages greater than 20 years.

Scenario 3 (initial nitrate concentration equal 2002-2004 nitrate distribution) indicated that if the NO$_3$ loading (only from residual nitrate) were to persist, the NO$_3$ concentrations within the aquifer would decrease within a few years. However, since Scenario 2 indicates that these concentrations do not account for what is currently observed in the aquifer, it is unlikely that they will provide realistic predictions of the future concentrations. A complicating factor, as discussed above, is that isotopic ages indicate that some of the water within the aquifer is more than 20 years old, suggesting that it may take longer than 10 years for the observed concentrations to decrease in the whole aquifer.
6.5 General Conclusions

There are many factors that control the distribution of NO₃ concentrations in the Abbotsford-Sumas aquifer. This research investigated heterogeneity present within the sand and gravel glacial outwash deposits that comprise the aquifer. Based on the lithology logs, geophysical surveys and the gravel pit photographs, it appears that different scales of heterogeneity exist. The well lithology logs used to create the regional groundwater flow model identified heterogeneity present on the scale of 100’s of metres to kilometres. The geophysical surveys, conducted as part of this research, identified layering consisting of fining and coarsening upwards sequences several metres thick and extending 10’s of metres. The smallest scale heterogeneity, present as layering on the scale of a few to 10’s of centimetres, was observed visually in the gravel pit.

Representing these various scales of heterogeneity within a numerical groundwater flow model largely depends on the scale of the model. Heterogeneity was best represented in the local scale model (approximately 20 km²) using vertical anisotropy, although travel times are perhaps underestimated based on isotopic ages of groundwaters. The implication of this, is that in the nitrate transport modelling conducted, it is likely that transport times are greater than observed.

Another important factor controlling NO₃ concentrations in the aquifer is the amount of NO₃ loading from agricultural activities above the aquifer. The concentrations reaching the aquifer from residual nitrate, as suggested by current BMPs, are clearly below those that are required to produce the observed NO₃ concentrations. Thus, NO₃ loading is likely higher than currently thought.

BMPs have changed in the last decade, and this should be leading to a decrease in the NO₃ concentrations in the aquifer, as is suggested by Hii et al. (in draft). However, the recent decrease may be due to a decrease in water levels in the aquifer. Isotopic ages greater than 10 years may suggest that decreases from the change in BMPs may still be seen in the future.
REFERENCE LIST


BC MAL. 2006. GIS Land use data for the Abbotsford aquifer.


City of Abbotsford. 2006. GIS Cadastral data.

Clague, J., 2006. Personal Communication. Department of Earth Sciences, Simon Fraser University, Burnaby, BC.


APPENDICES

Note: Appendix I is included in electronic format on the attached CD.

Appendix I: Abbotsford-Sumas aquifer: Compilation of a groundwater chemistry database with analysis of temporal variations and spatial distributions of nitrate contamination


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This is an PFD file named “Abbotsford-Sumas Nitrate Report” and is 2.40 MB.

Appendix I-A: Groundwater chemistry data in the Abbotsford-Sumas aquifer.

This is an MSExcel file named “Appendix I-A.xls” and is 4.14 MB.

Appendix I-B: Nitrate data in the Abbotsford-Sumas aquifer

This is an MSExcel file named “Appendix I-A.xls” and is 1.10 MB.

Appendix I-C: Numerical groundwater flow model datafile for nitrate data in the Abbotsford-Sumas aquifer

This is an MSExcel file named “Appendix I-A.xls” and is 191 KB.

Appendix II: Ground Penetrating Radar

Background on the technique and all collected data.

Appendix III: Borehole Logging

Background on the technique and all collected data.

Appendix IV: Lithology Logs

Lithology logs from the local model area.

Appendix V: Grain Size Analysis

Grain size analysis data from the PARC site.
APPENDIX II: GROUND PENETRATING RADAR

II.1 Applications of Ground Penetrating Radar

GPR has been used in a wide variety of environmental and earth science applications. Early uses of GPR included studies of densities and fractures of tunnel rocks (Cook, 1975), glaciers (Harrison, 1970; Watts and England, 1976), and salt domes (Holser et al., 1972), and to located buried pipes and cables (Annan et al., 1984). GPR has been applied successfully in different depositional environments, such as aeolian sediments (Bristow et al., 1996, Bristow, 1995), fluvial systems (Bridge et al., 1995; Tronicke et al., 2002; Young and Sun, 1999), lacustrine deltas (Jol and Smith, 1991), coastal systems (Engels and Roberts, 2005, Jol et al., 2002)) and glaciofluvial deposits (Huggenberger, 1993; Beres et al., 1995, 1999; Aiken, 1993; Beres and Haeni, 1991; Jakobsen and Overgaard, 2002; Olsen and Andreasen, 1995; Cassidy et al., 2003). GPR has also been used in many hydrogeological studies. Some of these include mapping of clay-sand boundaries (Young, 1995), the top of the saturated zone (Greenhouse et al., 1987; Bevan et al., 2003, van Overmeeren, 1994), imaging sedimentary units (Aiken, 1993; Greenhouse et al., 1987; Beres and Haeni, 1991; Jol and Smith, 1991; Rea and Knight, 1994; Beres et al., 1995, 1999; van Overmeeren, 1998; Asprion and Aigner, 1997; Olsen and Andreasen, 1995) and determining aquifer heterogeneity (Aiken, 1993; Gawthorpe et al, 1993; Rea and Knight, 1996).

II.2 Previous Work Using GPR for Aquifer Characterization

GPR has been used successfully to improve aquifer characterization at a number of sites. For example, Lunt et al. (2004) used GPR to image the Sagavanirktok River deposits in Alaska and were able to identify numerous features typical of Quaternary glaciofluvial outwash deposits.
Close et al. (2004) examined the presence of preferential flow paths at two locations in New Zealand, which contained heterogeneous aquifers. Preferential pathways were identified within these aquifers by correlating the GPR results to core logging.

Locally (within the Fraser Valley), an extensive GPR survey was conducted in the nearby Brookswood Aquifer, BC, which is a similarly heterogeneous aquifer comprised of Sumas Drift (Rea, 1996; Rea and Knight, 1998; Rea et al., 1994; Rea and Knight, 2000). The aquifer characterization was conducted by identifying and mapping radar architectural elements (Rea, 1996; Rea and Knight, 2000). These architectural elements were identified as hydrogeological units and were combined with drillers’ logs to reconstruct the depositional environment. The depth of penetration on the GPR sections ranged from about 5 m in the eastern part of the aquifer to about 15 m in the western part. The highly heterogeneous nature of the Brookswood aquifer was apparent in the GPR data, especially the presence of electrically conductive clay bodies, which were suggested to be no-flow boundaries. As part of that study, a single 20 m long GPR line was collected at the PARC site, which showed that there were several horizontal and slightly dipping reflectors visible within the top 6m of the profile.

Within the study region, Pullan et al. (2000) completed a single shallow seismic-reflection survey line along Boundary Road overlying the Abbotsford-Sumas aquifer. The quality of the data was considered to be poor due primarily to dry coarse-grained surface materials.

Irving and Knight (2003) conducted a cross-hole GPR survey between two piezometers at the PARC facility. This study examined radar wave velocity anisotropy in the saturated and unsaturated zone (horizontal velocity vs. vertical velocity). Their findings indicate that in the vadose zone, anisotropy is more significant than in the saturated zone since the coarse layers drain more readily than the finer layers. They determined an average vertical radar velocity of 0.12 m/ns in the vadose zone that is similar to the value of 0.118 m/ns determined in our study.
The cross-borehole survey was conducted in late July/early August, at the end of the dry summer season (Irving, personal communication, 2006). The ground has the lowest water saturation at this time of year, which may lead to an increase in the observed radar velocities compared to other times of year when the ground is more saturated.

II.3 Ground Penetrating Radar Theory

Ground-penetrating radar (GPR) is a geophysical method that provides non-destructive imaging of sub-surface objects (Davis and Annan, 1989). GPR systems emit radio-frequency electromagnetic pulses from the transmitting antenna into the subsurface. Propagation of the radar pulse depends on the electrical properties of the material it encounters. Changes in electrical properties result in both reflection and refraction of the pulse, or wavelet, in accordance with the principles of optics (Hanninen, et al., 1992). Finally, the reflected energy is recorded by the receiving antenna. Inhomogeneities in electrical properties are present in most hydrogeologic settings and are determined primarily by water content, dissolved minerals, grain size, and expansive clay and heavy-mineral content in the subsurface (Beres and Haeni, 1991).

The two electrical properties that affect radar transmission are the dielectric permittivity and the electrical conductivity. The relative dielectric permittivity (often referred to as dielectric constant) is a measure of the polarizability of a material when an electric field is applied to it relative to the polarizability of a vacuum (Sheriff, 1984). Electrical conductivity is a measure of the material's ability to conduct electricity. In materials with a low electrical conductivity (such as sand or gravel), the dielectric permittivity of the material dominates and can be related to the radar wave velocity of the material at different frequencies. Some typical dielectric permittivity values are provided in Table II-1. As the electrical conductivity of the material increases (as in silts and clays), so too does the attenuation of the radar wave.
Table II.1: Dielectric permittivity, radar velocity and electrical conductivity of some common geological sediments at 100 MHz (Data from Reynolds, 1997, and Davis and Annan, 1989).

<table>
<thead>
<tr>
<th>Material</th>
<th>Dielectric Permittivity</th>
<th>Radar velocity (m/ns)</th>
<th>Conductivity (mS/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air</td>
<td>1</td>
<td>0.30</td>
<td>0</td>
</tr>
<tr>
<td>Fresh water</td>
<td>81</td>
<td>0.033</td>
<td>0.5</td>
</tr>
<tr>
<td>Sea water</td>
<td>81</td>
<td>0.033</td>
<td>3x10^4</td>
</tr>
<tr>
<td>Dry sand</td>
<td>4 to 6</td>
<td>0.15</td>
<td>0.01</td>
</tr>
<tr>
<td>Saturated sand</td>
<td>30</td>
<td>0.06</td>
<td>0.1-1.0</td>
</tr>
<tr>
<td>Saturated silt</td>
<td>10</td>
<td>0.07</td>
<td>1-100</td>
</tr>
<tr>
<td>Saturated clay</td>
<td>8 to 12</td>
<td>0.06</td>
<td>2-1000</td>
</tr>
</tbody>
</table>

In geologic materials, radar wave reflections generally occur because of a change in lithology or volumetric water content (Figures II-1 and II-2) (Neal and Roberts, 2000). The larger the change in properties between materials, the stronger the reflection of the radar wave.

![Diagram](image_url)

Figure II.1: Schematic illustration of common-offset single fold profiling.
The maximum depth that can be imaged using GPR, or the radar range, depends on several factors. The most important are the GPR system performance, the attenuation in the ground, and the reflection properties at the interface where electrical properties vary (Davis and Annan, 1989). The principle limiting factor to the depth of penetration is the attenuation of the radar wave. Attenuation results primarily from spherical spreading losses and exponential material losses (Beres and Haeni, 1991). Scattering from boulders or other such features can also increase the effective attenuation.

Although the material properties are uncontrollable, there is some control that can be applied to the GPR system itself. Selection of antenna frequency can be an important factor for penetration depth, and typically GPR antennas range in centre frequency between 10 and 2000 MHz. Higher frequency antennas have higher resolution, but this is traded off with a lower depth of penetration. On the other hand, the lower resolution of lower frequency antennas can result in advantages from a lower degree of scattering: lower scattering attenuation and a simpler image to interpret due to the presence of fewer diffractions from scattering.
Previous studies have shown that in clay-free sand and gravel deposits, low-frequency radar waves can penetrate depths up to 90 ft (27 m) (Olhoeft, 1986). In materials that have high clay content, depth of penetration can be less than 3 ft (0.9 m) (Olhoeft, 1986).

II.4 Standard Survey Designs and Data Processing

There are four standard types of survey that can be conducted using GPR (Davis and Annan, 1989; Annan, 2001). The first, and most common, is constant-offset reflection profiling. The radar antennas are set in a fixed-offset configuration and are moved along the ground in a straight line. This produces an image of the two-way travel time to reflectors vs. antenna position. Radar wave velocity variation within the ground must be determined from either borehole logs, CMP soundings (described below), or most commonly from moveout velocity hyperbola fitting of diffractions observed in profile.

The second type of GPR survey that is typically done is the common-midpoint (CMP) sounding. CMP soundings are acquired by varying the spacing of the antennas around a central fixed location. CMP surveys are used to obtain an estimate of the radar signal velocity versus depth by measuring the change of the two-way travel time to the reflections as a function of antenna offset, i.e. measuring reflection moveout which is hyperbolic shaped (Annan, 2001). Radar velocity at the surface can be determined from the simple linear moveout of the direct ground wave using the simple equation, velocity = distance/time. Other, more complex equations are used to determine the velocities of subsurface reflectors from their hyperbolic moveout.

Similar to the CMP sounding, is the wide angle reflection and refraction (WARR) sounding. In a WARR sounding, one antenna is held fixed while the other is moved away. This method was common in the early years of GPR work, but is now rarely used (Annan, 2001).

The final type of survey is called transillumination (such as in cross-hole surveys) and is used to “look” through the material (Davis and Annan, 1989). This application usually only
occurs in mines, between boreholes or in structural integrity tests. One antenna is fixed at the side of a pillar or wall or within a borehole. The other antenna moves past the first antenna on the other side of the wall or pillar, or in and adjacent borehole.

The basic processing that is applied to GPR data is described in detail in Annan (2001). If necessary, the first step is time-zero correction. GPR data always requires "dewow" filtering, basically temporal filtering to remove a bias in the trace zero-level that decays with time down the trace. Usually after any type of filtering a time varying gain is applied, which equalizes the amplitude of the signal to account for the rapid attenuation of signal strength with depth. At this stage, depending on necessity, additional processing may be applied of which there are many different types of temporal and spatial filters that can further refine the radar image such as bandpass filtering and migration.

II.5 Field Methodology

The radar data were collected using a Sensors & Software pulseEKKO 100 ground-penetrating radar unit with 50 MHz and 100 MHz antennas. This GPR unit consists of transmitting and receiving antennas, a control box and a laptop.

Both common-offset and common-midpoint data were collected as part of the GPR survey. A common-midpoint survey was conducted first along Line 1 at both 50 MHz and 100 MHz antenna frequencies. The initial separation between the antennas was 0.2 m, with an increase in separation of 0.2 m for each subsequent set of measurements.

Common-offset surveys were conducted at both 50 MHz and 100 MHz antenna frequencies on 7 lines at the PARC site (Figure II-3). The lines formed a grid with line spacing varying between 20 to 190 m. Profile surveys at 100 MHz used a constant antenna offset of 1 m (based on the results of the CMP survey) with a station spacing of 0.2 m. Profile surveys at 50
MHz used a constant offset of 2 m with a station spacing of 0.4 m. Figure II-4 shows the GPR survey being completed at the PARC site.

Figure II.3: Location of GPR lines at the PARC site. The black square indicates the location of the centre of the CMP soundings.
II.6 Data Processing

GPR data were processed using the software ReflexW (Sandmeier, 2005). For the paper in Chapter 3, the data were reprocessed using a different time zero correction resulting in the velocities indicated in italics, and an energy decay gaining. The processing steps are outlined below;

1. Time zero correction;
2. Dewow filtering (moving average windows of 30 ns and 15 ns for 50 MHz and 100 MHz data, respectively);
3. Gaining (automatic gain control (100 ns window length);
4. Velocity Analysis (diffraction hyperbola fitting);
5. Constant Velocity Migration (at 0.08 m/ns (0.100 m/ns));
6. Regaining and Depth Scale (using constant average velocity of 0.105 m/ns (0.118 m/ns)).

After applying a constant time-zero shift, a moving average dewow filter with a window length of 30 ns for the 50 MHz antenna data, and 15 ns for the 100 MHz antenna data was applied. This was followed by an automatic gain control with a 100 ns window length. All of the profiles exhibited prominent near surface diffractions (from boulders) that interfered with the stratigraphic image. Velocities determined from the diffraction hyperbolas ranged from 0.075 – 0.085 m/ns (0.095 – 0.105 m/ns). These diffractions were therefore filtered by applying a constant velocity migration to the data using a velocity of 0.080 m/ns (0.100 m/ns). There was a strong reflection from the water table at the PARC site, and the velocity determined from the known depth of the water table at the time of the surveys provided a slightly higher velocity of 0.105 m/ns (0.118 m/ns). The depth scale on the profiles was calculated using this higher velocity.

The difference between the velocities determined from the diffraction migration and those from the water table produces some uncertainty. It is unusual for radar velocities to increase with depth. This may be explained at the PARC site because the majority of the diffractions that were used for the migration were located in the top few metres below the ground surface, which may be more organically enriched soil and more saturated due to irrigation, resulting in an increased velocity. In this case, with a highly saturated near surface and a large vadose zone, it is realistic to expect the velocity to increase towards the water table. Below the water table, the velocity would be expected to decrease again because of the saturation of the material.
II.7 CMP Results

The CMP soundings that were collected along Line 1 at the PARC site are shown in Figures II-5 and II-6. As there are no strong reflectors visible in the subsurface, only the near surface velocity was determined from the first arrival of the ground wave. This is indicated with a red dotted line on the sections. By using the equation velocity=distance/time, it was determined that for both the 50MHz and the 100Mhz antenna frequencies that the direct radar wave velocity was 0.105 m/ns.
Figure II.5: CMP sounding on Line 1 using 50 MHZ antennas. The top figure is the unprocessed data, and the bottom figure shows the processed data. The dotted line indicates the first arrival of the ground wave.
Figure II.6: CMP sounding on Line 1 using 100 MHz antennas. The top figure is the unprocessed data, and the bottom figure shows the processed data. The dotted line indicates the first arrival of the ground wave.
II.8 GPR Sections

The GPR reflection sections that were collected at each of the seven lines at the PARC site are shown in Figures II-7 through II-20. Each figure represents either the 50 MHz or 100 MHz data as indicated in the figure caption. The top image in each figure is the raw, unprocessed data for each line. The bottom image is the same data after processing. The depth scale on the right side of the images is based on the direct radar wave velocity of 0.105 m/ns. The 50 MHz profiles show a deeper penetration than the 100 MHz surveys, however, the 100 MHz surveys have a significant increase in resolution above the water table. The water table is represented as a strong reflector between 360 and 400 ns on most of the sections. Below this, especially in the 100 MHz, the signal is often lost due to increased attenuation with the more conductive water.

The interpretation is based on reflection configurations described in Beres and Haeni (1991). All of the GPR sections show the same types of reflectors. These are generally undulating to chaotic in nature, with diffractions occurring predominantly close to the ground surface. The undulating character is dominant suggesting that formation is comprised of bedded sands and gravels. In areas that appear more chaotic in nature, smaller scale (not fully resolved) sand and gravel cross-bedding and/or smaller lenses are likely present, with larger diffractions caused by boulders.
Figure 11.7: Line 1 Reflection survey at 50 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.8: Line 1 Reflection survey at 100 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.9: Line 2 Reflection survey at 50 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.10: Line 2 Reflection survey at 100 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure 11.11: Line 3 Reflection survey at 50 MHz. The top figure shows the raw, unprocessed data, and the bottom figure shows the data after processing.
Figure II.12: Line 3 Reflection survey at 100 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure 11.13: Line 4N Reflection survey at 50 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.14: Line 4N Reflection survey at 100 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.15: Line 4S Reflection survey at 50 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.16: Line 4S Reflection survey at 100 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.17: Line 5 Reflection survey at 50 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.18: Line 5 Reflection survey at 100 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.19: Line 6 Reflection survey at 50 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
Figure II.20: Line 6 Reflection survey at 100 Mhz. The top figure shows the raw, unprocessed data, and the bottom figure show the data after processing.
APPENDIX III: BOREHOLE LOGGING

III.1 Borehole Logging Theory

Borehole geophysics is defined as “the science of recording and analyzing continuous or point measurements of physical properties made in wells or testholes” (Keys, 1989). These measurements are made by lowering a probe into the borehole and electronically transmitting the data to a recorder at surface, commonly a laptop, where the data are recorded as a function of depth. These measurements are usually referred to as borehole logs and there are many types of logs that can be conducted. Common logs include:

- Caliper logs - provide a continuous record of the hole diameter.

- Natural-gamma logs - record the amount of natural gamma radiation emitted by the material surrounding the borehole due to trace elements of potassium uranium and thorium.

- Single-point resistance logs - record the electrical resistance between an electrode in the borehole and an electrode at the ground surface.

- Spontaneous potential logs - record voltages that develop between the borehole fluid and the formation fluids.

- Normal-resistivity logs - record the electrical resistivity of the surrounding material over different ranges from the borehole by using different spacings of potential electrodes on the probe.
Electromagnetic-induction logs (also known as conductivity logs) - record the conductivity (inverse of resistivity) of the material by electromagnetically inducing a current to flow in the material.

Fluid-resistivity logs - record the electric resistivity of the water in the borehole.

Fluid temperature logs - record the temperature of the water in the borehole.

Flowmeter logs - record the rate and direction of fluid movement in a borehole (with or without pumping).

Television logs record a color (or black and white) video of the borehole.

Optical and acoustic-televiewers log oriented digital scans of visual and acoustic reflectivity of the borehole wall, respectively. (Keys, 1997, 1989)

During this study, natural-gamma and EM-induction (conductivity) logs were collected from the piezometers at the PARC site. Further description of these methods is provided below.

III.1.1 Natural-Gamma

Gamma rays are produced by the natural decay of radioactive isotopes. The most significant radioactive isotopes that occur naturally in unconsolidated sediments are potassium-40, and the daughter products of uranium and thorium decay (Keys, 1997). Shales and clays have higher concentrations of these radioactive isotopes than other rock and sediment types, although exceptions can occur (Douma et al., 1999).

The natural gamma tool records the total radiation produced by the lithologic formation as well as any artificial radioactivity that might be present. Artificial radiation can be identified only through spectral gamma logging. Gamma radiation is measured from the geologic material immediately surrounding the borehole, to a maximum distance of approximately 30cm from the tool (Douma et al., 1999). The tool uses a sodium iodide scintillation detector coupled to a
photomultiplier. Gamma rays are released from the formation and, consequently, electrons are produced in the scintillation crystal through one of three basic gamma ray interaction mechanisms, and the electrons are detected by the photomultiplier (Ellis, 1987).

III.1.2 Electromagnetic Induction (Conductivity)

The electrical conductivity of a porous unconsolidated material is a function of the combined electrical conductivity of the matrix and the pore fluid. Conductivity is the current carrying capacity of a material (Rider, 1986). The conductivity probe induces a current in the formation around the borehole (within approximately 0.75 m of the probe) via an electromagnetic transmitter coil, and the resulting electromagnetic response is measured by the receiver coil. The quadrature component of the measured field is directly proportional to electrical conductivity. If the pore fluid conductivity is low (air or fresh water) then the bulk conductivity of the material mainly reflects that of the matrix material. Coarse grained materials generally have low conductivities, while fine grained materials (especially clay) have higher conductivities. If the pore fluid conductivity is high (saline water), then the differences between the coarse and fine grained materials will be less noticeable (Douma et al., 1999).

The conductivity tool works by an AC current applied to the transmitter coil which, in turn, causes an alternating electromagnetic field (the primary field) to be transmitted into the surrounding formation. The time varying primary magnetic field, in turn, induces a flow of alternating eddy currents in the surrounding formation that try to counteract the change in the magnetic field. These eddy currents produce secondary magnetic fields, which induce a current in the receiving coil in the tool (Keys, 1997). The magnitude of the current in the receiving coil (specifically, the quadrature component of receiver current, i.e. the component that is 90° out of phase with the transmitted field) is proportional to formation electrical conductivity (Douma et al., 1999).
III.2 Previous Work Using Borehole Logging for Aquifer Characterization

Borehole logging is a group of methods that have been used since the 1930's to investigate the subsurface (Segesman, 1980). Borehole logging is now used in a wide range of exploration, engineering, and environmental/groundwater applications. Both natural-gamma and conductivity logging are widely used in environmental/groundwater applications (Keys et al., 1993).

Much of the work that uses borehole geophysics for aquifer characterization is for examining aquifer lithology and improving stratigraphic correlation, and water quality. Aquifer lithology is commonly examined using both natural-gamma and conductivity logging (Cromwell, 1992; Nobes and Schneider, 1996; Barrash and Morin, 1997; Pullan et al., 2002; Siron and Segall, 1997; Paillet and Reese, 2000). Because natural-gamma is sensitive to clay content and, therefore, grain size, in Quaternary deposits it is often the only logging technique used to examine aquifer lithology (West 2002; Norris, 1972; Baldwin and Miller, 1979; Dixon-Warren and Stohr, 2003). Conductivity logs, which measure changes in the conductive properties of the material as well as the groundwater, can also be used to examine groundwater quality (Alger and Harrison, 1989; Keys, 1989).

III.3 Previous Work Using Borehole Geophysics in the Abbotsford-Sumas Aquifer

In 1993, as part of the Fraser Lowland Hydrogeology Project undertaken by the GSC, two piezometers at the PARC site (92-3 and 92-4) were logged using natural-gamma, conductivity and magnetic susceptibility sondes (GSC, 2003). The GSC natural-gamma and conductivity logs are included with our corresponding logs in the results section.

Irving and Knight (2003) conducted cross-hole GPR logging between two piezometers at the PARC site in order to investigate saturation-dependent radar wave velocity anisotropy in the saturated and unsaturated zone. Their findings indicate that in the vadose zone, anisotropy is
more significant than in the saturated zone since the coarse grained layers drain more readily than
the finer grained layers. They determined an average vertical radar velocity of 0.12 m/ns in the
vadose zone that is similar to the value of 0.118 m/ns determined in our study. They completed
additional cross-hole GPR logs at PARC in 2004, which are still in the process of being analyzed
(Irving, personal communication, 2006).

III.4 Standard Borehole Logging Methods and Data Processing

Once the type of geophysical logs that is desired has been decided, the appropriate probe
is attached to the electric cable by a cable head screwed onto the top of the probe. The cable is
used to transmit recorded data to surface. The winch lowers and raises the probe within the
borehole (Figure III-1). Within the winch unit, there is a speedometer, which measures the
logging speed and probe depth. Logging speed is an important variable because it may affect
logging data quality and resolution. Logging speeds can range from less than 1 to 6 m/min (Ellis,
1987). Slower logging speeds usually result in greater detail and in better counting statistics for
nuclear probes (e.g., natural-gamma). On the other hand, faster logging speeds lead to less tool
sticking that can adversely affect all logs. (Keys, 1989; 1997; Ellis, 1987).

Figure III.1: Typical setup for borehole logging.
If calibration of the probe is required, this is completed before the data are collected. Logging can occur in either an up or down direction within the borehole, but is usually done in an up direction to avoid the probe sticking within the borehole. The measurements made by the probe are recorded and displayed on a laptop computer in real time.

Minimal data processing is usually required for borehole logs. Other than simple depth corrections, the two most common types of processing are smoothing and/or some other form of filtering. Smoothing reduces the scatter of the log data, especially in nuclear logging.

III.5 Field Methodology

The borehole logs were collected using a Mount Sopris MGX-I1 portable digital logger, including a 305m winch. Two tools were used: a 2PGA-1000 natural gamma tool and a Geonics 2PIA-1000 electromagnetic induction tool. The field set up is shown in Figure III-2.

![Borehole logging setup at PARC site. Inset: 2PGA-1000 natural gamma tool.](image-url)
Natural gamma logs were completed for all piezometers at the site except for CDA2 (see Figure III-3 for piezometer locations). The logging speed was 2.75m/min, resulting in a sampling interval of 0.005m. Logs were collected in an upwards direction.

Electrical conductivity logs were completed for all piezometers at the site except for CDA2 (See Figure III-3 for piezometer locations). The conductivity logs were collected at 2.75m/min, resulting in a sampling interval of 0.01m. Logs were collected in an upwards direction.

Figure III.3: Location of piezometers logged at the PARC site.

III.6 Data Processing

Processing was only completed for the natural-gamma data. The data were smoothed using a 21-point moving average filter to reduced noise in the data. No processing was required for the conductivity data. The data were displayed using the software LogView (Markarian et al., 1995).
III.7 Results

Natural gamma borehole logs (both unsmoothed and smoothed) are presented along with electrical conductivity logs in Figures III-4 through III-12. The GSC (2003) logs of piezometers 94-2 and 94-3 are also shown in Figures III-10 and III-11. In the natural-gamma logs, fining upwards sequences can be seen in the sand and gravel. This would suggest a gradual change in the energy of the deposition environment resulting in a gradual change in grain size. There are also sudden shifts in the gamma count (either up or down), most or all of which are probably erosion surfaces or scour and fill features. The duplicate logs for 94-2 and 94-3 show a similar gamma response.

There appears to have been a calibration problem within the logging tool, which resulted in the conductivity logs being shifted to lower conductivity by an unknown number. Therefore, only the relative changes in conductivity could be examined. However, when these logs are compared with the GSC data in boreholes 94-2 and 94-3 (Figures III-10 and III-11), they show similar responses to the formation, and if acquired under similar ground conditions, the GSC logs could be used to derive an approximate correction factor. The amplitude of the observed changes is similar, suggesting that the relative changes in conductivity in our logs can still be used for interpretation. In all of the boreholes, there is an increase in the conductivity at the water table, indicating that the formation water has an effect on the conductivity.

In the piezometers at the PARC site, the relationship between the recorded natural-gamma and conductivity values provides information about the pore fluid surrounding the piezometers. A plot of gamma vs. conductivity for piezometer 91-4 is shown in Figure III-13. In the unsaturated zone (oval enclosure on Figure III-13), there appears to be no relationship between the two values. In the transitional zone, the conductivity ranges between unsaturated and saturated values, while the gamma values remain as they were in the unsaturated zone. This transitional zone likely represents the capillary fringe above the water table. In the saturated zone
(between the lines), an inverse relationship between the conductivity and gamma is observed, although there is some scatter in the gamma values. This inverse relationship indicates that the observed increases and decreases in the conductivity are related to changes in pore fluid conductivity rather than changes in grain size or clay content. This increased conductivity in saturated, slightly, coarser-grained material may be partly—or entirely—due to greater mobility of ions in larger/better connected pore spaces, rather than increased ion concentration. Determining the balance of these two roles (greater ion concentration, vs. greater mobility) would require targeted groundwater sampling within zones of higher and lower gamma counts to determine the respective groundwater chemistry; unfortunately, this is not possible with the existing piezometer placements.

There is always some inherent error associated with the gamma counts recorded in the logs. Piezometer 91-4 was logged in both the upwards direction (as discussed above) and the downwards direction (Figure III-14). The logging speed was 1.5 m/min, resulting in a sampling interval of 0.05m. This was a slower logging speed than the logging done in the upwards direction. Figure III-14 shows that there is little difference between the logs in both directions, reducing the uncertainty associated with the interpretations of the logs. As well, the correlation of units across several boreholes (Figure 3.6) would suggest that there is minimal effect from the borehole itself.
Figure III.4: Geophysical logs recorded in piezometer 91-1. The water table in the piezometer is indicated.
Figure III.5: Geophysical logs recorded in piezometer 91-2. The water table in the piezometer is indicated.
Figure III.6: Geophysical logs recorded in piezometer 91-3. The water table in the piezometer is indicated.
Figure III.7: Geophysical logs recorded in piezometer 91-4. The water table in the piezometer is indicated.
Figure 11.8: Geophysical logs recorded in piezometer 91-5. The water table in the piezometer is indicated.
Figure III.9: Geophysical logs recorded in piezometer 91-7. The water table in the piezometer is indicated.
Figure III.10: Geophysical logs recorded in piezometer 94-2. The red line is the data collected by the GSC (2003). The water table in the piezometer is indicated.
Figure III.11: Geophysical logs recorded in piezometer 94-3. The red line is the data collected by the GSC (2003). The water table in the piezometer is indicated.
Figure III.12: Geophysical logs recorded in piezometer CDA1. The water table in the piezometer is indicated.
Figure III.13: Gamma vs. Conductivity plot in piezometer 91-4. Values from the unsaturated zone are generally grouped within the oval, while the values from the saturated zone have a higher conductivity and lower conductivity.
Figure III.14: Comparison of upward and downward logs for natural gamma in Piezometer 91-4.
## APPENDIX IV: LITHOLOGY LOGS

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APPENDIX V: GRAIN SIZE ANALYSIS

Figure V.1: Grain size distribution in piezometer 91-1.

Figure V.2: Mean gamma count versus grain size for piezometer 91-1.
Figure V.3: Grain size distribution in piezometer 91-2.

Figure V.4: Mean gamma count versus grain size for piezometer 91-2
Figure V.5: Grain size distribution in piezometer 91-3.

Figure V.6: Mean gamma count versus grain size for piezometer 91-3.
Figure V.7: Grain size distribution in piezometer 91-4.

Figure V.8: Mean gamma count versus grain size for piezometer 91-4.
Figure V.9: Grain size distribution in piezometer 91-5.

Figure V.10: Mean gamma count versus grain size for piezometer 91-5.
Figure V.11: Grain size distribution in piezometer 91-7.

Figure V.12: Mean gamma count versus grain size for piezometer 91-7.