THE ROLE OF AQUIFER HETEROGENEITY IN SALTWATER INTRUSION MODELING, SATURNA ISLAND, B.C., CANADA

By

Emilia Liteanu
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APPROVAL

Name: Emilia Liteanu

Degree: Master of Science

Title of Thesis: The Role of Aquifer Heterogeneity in Saltwater Intrusion Modelling, Saturna Island, British Columbia

Examining Committee:

Chair: Dr. Peter Mustard
Associate Professor

Dr. Diana Allen
Senior Supervisor
Associate Professor

Dr. John Clague
Supervisory Committee Member
Professor

Dr. Doug Stead
Supervisory Committee Member
Professor

Mr. Mike Wei
External Examiner
Ministry of Water, Land and Air Protection
Senior Groundwater Hydrologist

Date Approved: November 27, 2003
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Author: Emilia Liteanu

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(Name)
ABSTRACT

In coastal aquifers, presence of heterogeneity introduces complexity to the configuration of the saltwater - freshwater interface. Field studies, in combination with pumping test analyses and numerical simulations conducted as part of this thesis, provide insight into the complexity of groundwater flow on the Gulf Islands, Canada.

The geology of the Gulf Islands consists of an alternating sequence of sandstone- and mudstone-dominant formations, which are extensively fractured. Pumping tests conducted throughout the region provided data that were analyzed using various analytical methods to determine a range of hydraulic parameters that characterize the geological formations. Most tests display a period of linear flow when the pumping well was located near a fracture zone or fault. The derivative method was used to identify the radial flow period, and radial flow data were analyzed to provide estimates of the hydraulic parameters. No specific trends were identified that might suggest lithology-dependent hydraulic parameters, despite field evidence showing a relation between permeability and lithology.

Density-dependent flow and solute transport simulations were carried out for Saturna Island using USGS SUTRA to model the position of the saltwater-freshwater interface under varying degrees of layered heterogeneity. Steady-state simulations, calibrated against groundwater chemical data, were undertaken to determine if the hydraulic parameters used in the model are consistent with those determined directly from pumping tests. Models simulations indicate that the magnitude of the permeability and the nature of layering exercise a major control on the magnitude and appearance of the freshwater-saltwater interface, and that the permeability values needed are roughly one order of magnitude higher than calculated.

Transient simulations were conducted to simulate the behaviour of the freshwater-saltwater interface over the last 12,000 years, and thereby further constrain the hydraulic parameters used in the model. Simulations show that the current configuration of the saltwater-freshwater interface is consistent with that predicted from the late Pleistocene sea level history, which included a brief period of island submergence roughly 12,000 years ago, followed by rapid rebound and stabilization of sea level at its current elevation.
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CHAPTER 1 - INTRODUCTION

Background

The study of coastal and island aquifers is of critical importance because of growing concern in many regions around the world about groundwater depletion and the associated contamination from encroaching seawater. Worldwide, groundwater represents 0.6% of the total water resources on Earth. However, because the majority of surface water is saltwater (97%), groundwater is the main source of potable water (Bear and Cheng, 1999). As a result of increasing population, global freshwater consumption has increased significantly. In coastal regions, the depletion of fresh groundwater resources is complicated by contamination with salt. Therefore, protection and management of groundwater resources is a growing concern worldwide.

Coastal aquifers exist where large land areas are in direct contact with seawater, or on oceanic islands which are surrounded by seawater. Because seawater forms a physical hydraulic and chemical boundary to these aquifers, there is an interaction between freshwater discharging to the coast, and seawater. At depth, there is an interface between the two water types, which is either abrupt or transitional in character. Under natural "equilibrium" conditions, groundwater discharges towards the sea or ocean, and freshwater flows above the saltwater (Figure 1.1).

The groundwater flux is controlled by the hydraulic gradient. Chemical transport of solutes occurs by a combination of advection, whereby solutes are transported along with the moving water, and molecular diffusion. The width (sharp or transitional) and position of the freshwater-saltwater interface depend on the magnitude of the gradient.

Saltwater intrusion refers to the movement of salty water into freshwater aquifers, which results in contamination of freshwater resources. The magnitude of saltwater intrusion can be expressed as a function of chloride concentration. Concentrations higher than 150 mg/l Cl result in poor drinking water quality. Changes to either boundary conditions, namely sea level or the freshwater flux discharging to the coast, will result in a shift in the interface position. The boundary conditions are dynamic and, as a consequence, coastal aquifers are affected by both natural and
Figure 1.1 Movement of saltwater and freshwater in coastal areas
anthropogenic factors: sea level changes, waves, currents, storms, climate change (encompassing changes in precipitation and evapotranspiration), surface drainage and exploitation. Each of these factors affects the interaction between fresh groundwater and seawater. Figure 1.2 summarizes the factors affecting groundwater and the effects on fresh groundwater resources in coastal regions (summarized from Oude Essink, 2001).

**Relative Sea Level Changes**

As coastal aquifers are in direct contact with seawater, the hydrogeology is influenced by relative sea level changes. A significant rise in sea level may lead to a major decrease in the size of freshwater lens (Sherif and Singh, 1999), while a sea level regression will result in the saltwater-freshwater interface moving seaward. Storms, waves, and tides are few factors that affect groundwater conditions in coastal aquifers over the short-term.

**Changes in the Hydrological Regime**

Changes in the hydrological regime similarly influence coastal aquifers. These changes include primarily changes in precipitation and evapotranspiration, which together control the amount of recharge and the groundwater flux. A reduction in freshwater flux causes movement of saltwater-freshwater interface inland.

Studies conducted by Sherif and Singh (1999) and Jacoby (1990) indicated that climate change plays a key role in the water balance. Thus, studies of climate change impacts on aquifers can provide insight into the hydrogeology in the geologic past and into the future.

**Human Activities**

Groundwater consumption, beyond that replenished by recharge, leads to a reduction of groundwater flux. Not only does this reduction in flux cause the freshwater-saltwater interface to move inland, but it also results in over-exploitation of the resource.

**Paleo-Groundwater**

A complicating factor in many coastal regions is the presence of old or "paleo" groundwater, which is characteristically saline. Paleo-groundwater is water that has been entrapped in the aquifer and has remained there for a considerable time. Continual freshwater replenishment by recharge and/or exploitation may result in paleo-
Figure 1.2 Factors affecting coastal aquifers (modified after Oude Essink, 2001)
groundwater coming out of storage and contaminating fresh groundwater resources. In coastal aquifers, these waters may offer an additional source of salinity to that of saltwater originating from the ocean (Yecheli et al., 1996, Hahn et al., 1991).

Sustainable use of fresh groundwater resources can be realized by identifying, monitoring, studying and managing coastal aquifers. Management varies from area to area within the same country and also from country to country, and involves consideration of numerous factors. Management is often based on models that predict an accurate flow pattern, which show how the aquifer evolves over time under different scenarios. For effective management of fresh groundwater resources, it is essential to predict the extent of saltwater intrusion and the responses of the aquifer to different stresses. These predictions require an understanding of the hydrogeology of the region, and often, an accurate model that characterizes the groundwater regime in the aquifer.

The hydrogeology of homogeneous coastal aquifers has been well-studied in many areas including: Staten Island (Pucci, 1987) and Long Island in New York (Stum, 2001), but in few cases in heterogeneous aquifers (Simmons et al., 2001). Numerous studies have used analytical or numerical approaches in an attempt to model the position of the freshwater-saltwater interface under steady-state conditions (e.g., Egypt: Sherif and Al Rashed, 2000), or the dynamics of the interface under time-varying conditions (e.g., Belgium: van Camp and Walraevens, 1999). In addition, several studies have assessed the impact of such factors as increased water demand, reduced recharge, or climate change (e.g., The Netherlands: Oude Essink, 1999; USA: Mehta et al., 2001, Israel: Yakirevich et al., 1998; Sherif and Singh, 1998; and India: Bobba, 2000). However, most of these studies have been limited to simple aquifer geometries; specifically, homogenous aquifer properties. There are relatively few studies that have attempted to characterize and model heterogeneous coastal aquifers.

This M.Sc. thesis aims to examine the role of heterogeneity in an island aquifer setting by conducting a sensitivity analysis in a heterogeneous fractured aquifer, using SUTRA 2-D (USGS, 1984). The study area is the Canadian Gulf Islands, which are situated off the southwest coast of British Columbia, Canada. Specifically, Saturna Island will be used as a case study site. Saturna Island is the southernmost Gulf Island and lies between North 48'87" latitude and East 123'13" longitude on the southwest coast of British Columbia, Canada.
Previous Research on the Gulf Islands

Previous hydrogeological and geological studies on the Gulf Islands have included: 1) chemical analyses of groundwater, surface waters and spring waters, which were used to establish a possible geochemical evolution of groundwater on Saturna Island (Allen and Suchy, 2001); 2) stable isotope analyses ($\delta^{18}$O and $\delta^{34}$S in dissolved sulphate, and $\delta^{18}$O and $\delta^{2}$H in water) that were used to determine the possible origin of groundwater on the island (Allen, in press); 3) geophysical surveys (borehole and surface) to study the potential for preferential flow and saltwater intrusion near faults and fracture zones (Allen et al., 2002), 4) structural measurements of over 8000 fractures at 157 stations on 8 islands in the southern Gulf Islands region (Mackie, 2002); 5) hydraulic testing conducted on several islands (Abbey, 2000; Allen et al., 2003); and 6) documentation of the Pleistocene history and record of sea level change for the south British Columbia coast and Vancouver Island (James and Clague, 2001; Clague, 1986; Clague, 1983). Important conclusions from some of these studies and their relevance to the purpose of this research are presented below.

Hydrochemical Studies

Based on chemical analyses conducted on water samples from Saturna Island, Allen and Suchy (2001) concluded that the groundwater has evolved along two geochemical evolution paths as Figure 1.3 illustrates:

- **Path 1**: Characterized by cation exchange, whereby Ca-HCO$_3$ rich freshwater undergoes cation exchange (exchange for Na) and then mixes with Cl-rich water at depth.

- **Path 2**: Characterized by direct salinization whereby fresh groundwater mixes directly with saltwater.

This two step process results in mixing of freshwater with saline groundwater in the natural saltwater wedge at depth, and modern-day saltwater intrusion, respectively.
Figure 1.3 Piper diagram for waters collected in the East Point area of Saturna Island (data from Suchy, 1998).
**Structural Studies**

The hydrogeology of the Gulf Islands is directly related to the intensity of fracturing as well as the type of fracturing. The primary porosity of the Nanaimo Group is about 5% (as determined from oil industry wells), and therefore, fractures (if connected) in the form of secondary porosity offer the only substantial permeability. Measurements of fracture attributes and distribution, and their relation to faults and folds, along with geophysical methods of investigation revealed two main points. First, that fracture density increases with closer proximity to faults (Allen et al., 2002; Mackie, 2002). This is supported by drilling records, which indicate that wells placed near fracture zones have higher yields (e.g., 200 US gpm) compared to wells placed away from fracture zones (e.g., 5 US gpm). Second, fracture spacing (as bedding perpendicular joints) was observed to be smaller in the thinner mudstone interbeds compared to the thicker sandstone beds (Mackie, 2002), suggesting that mudstone-dominant units and mudstone interbeds within sandstone-dominant formations are more intensely fractured, and probably constitute dominant water-bearing aquifers on the islands compared to the thicker sandstone-dominant formations. Driller logs also indicate that groundwater is derived from the fractured zones that are often present at contacts with mudstone units, supporting these observations.

**Quaternary Studies**

During the late Pleistocene, the Gulf Islands were covered by a thick sheet of ice that depressed the land to about 150 meters below the current elevation. During ice sheet retreat, about 12,000 years ago, the Gulf Islands were briefly inundated by seawater, and very quickly thereafter, about 11,000 years ago, the area rebounded. Glacio-isostatic rebound was accompanied by a drop in sea level to slightly below its present position in most areas. Thus, between 11,000 and 10,000 years ago, the relative sea levels fell 30 meters below current levels (Clague, 1998).
Relevance to Current Research

Allen and Suchy (2001) proposed that the origin of Na is remnant seawater that was introduced (along with Cl) into the rock sequence during the late Pleistocene when the island was submerged. Isostatic rebound, following submergence, has subsequently removed Cl, by flushing the aquifer with fresh (HCO₃ rich) infiltrating groundwater. It is hypothesized that the Na largely remained in place because of sorption to clay minerals, but is now being released as fresh water continues to infiltrate. The current hydrochemical model for shallow to intermediate flow regimes would call for fresh groundwater and remnant Na-rich groundwater, while deeper flow systems, and groundwater discharging along the coast, would have higher concentrations of Cl.

However, for this model to be correct, there must be sufficient time for seawater to have entered the aquifer when the island was submerged, and following rebound, freshwater infiltration must have been sufficient to flush Cl from the aquifer and account for its current concentration.

The evolution of groundwater as described above is complicated by two main factors: 1) present day saltwater intrusion, and 2) aquifer heterogeneity. Local occurrences of saltwater intrusion are expected to be a consequence of reduction in the groundwater flux toward the coast under conditions of water table lowering associated with groundwater withdrawal or climate change (reduction in recharge). The occurrence of saltwater intrusion is generally restricted to low-lying areas adjacent to the coast, particularly, narrow peninsulas (Allen and Suchy, 2001). At a local scale, wells that are contaminated by sea water are randomly distributed due to the complexity of fracturing on the islands. Movement of groundwater at a regional scale is expected to be affected by the type of lithology (sandstone- or mudstone-dominant) and by the presence of major fracture zones and faults.

Purpose

The primary purpose of this research is to test the conceptual model for the geochemical evolution by modeling the dynamics of the saltwater-freshwater interface under the constraints of the post-glacial history of the region. Modeling the movement of the saltwater–freshwater interface over the last 12,000 years will provide a means to test the theory that a period of 1,000 years (between ice retreat and rebound) is sufficient to
result in at least partial saturation of the island with seawater, and to test if Cl has had sufficient time to be removed from shallow and intermediate aquifer depths and be at higher concentrations at depth near the coast. The chemical evolution of groundwater, therefore, is anticipated to be constrained by the late Pleistocene history of the region.

In order to develop a flow and transport model for Saturna Island, it is necessary to consider aquifer heterogeneity as represented by fracturing. Due to the complexity of the geology and structure of the islands, models will be constructed in cross-section. Consequently, the role of vertical fracture zones and faults will not be captured. Rather, the models will incorporate a layered form of heterogeneity (represented as Equivalent Porous Media) that is associated with the difference in fracture density of the lithologic units (i.e., mudstone- and sandstone-dominant formations). Heterogeneities in a hydrogeological system play an important role in controlling groundwater flow and solute transport at both local and regional scales. In variable density flow systems, heterogeneity in the hydraulic parameters can disturb the flow by generating instabilities over different scales, from the pore scale, to local or regional scale. Stratified systems are one type of environment that requires additional modeling studies to examine the behaviour of the saltwater-freshwater interface.

Scope of Work

1. To determine appropriate hydraulic parameters for each geologic formation based on previous aquifer tests conducted on Gulf Islands and by considering the placement of wells with respect to proximity to fracture zones, and to geologic formation.

2. To build a conceptual model that represents the hydrogeology of Saturna Island that can be used to construct a numerical model to specifically address the purpose of this research, namely modeling the dynamics of the freshwater-saltwater interface over the past 12,000 years.

3. To attempt to calibrate model to the chemical data acquired during previous research to verify the model against the Pleistocene history of the region.

4. To determine the role of heterogeneity, as represented by geologic layers, in solving for density dependent flow and solute transport in the vicinity of the freshwater-saltwater interface.
Methodology

The scope of this thesis involves determining the dynamics of freshwater-saltwater interface on Saturna Island from the late Pleistocene to present. Previous studies indicate that this can be done successfully using numerical modeling. The USGS SUTRA code is suitable for modeling density dependent flow and, consequently, will be used for this research.

Modeling also requires a conceptual model that describes physical and chemical controls on the system, and involves defining:

- Topographical, geological and structural features of Saturna Island;
- Hydraulic parameters (e.g., hydraulic conductivities and specific storage values obtained by analyzing available pumping test data);
- Boundary conditions that suit the hydrogeological setting, including approximate late Pleistocene to current sea level history;
- Estimates of recharge from historic climate data;
- An approximate position of current saltwater–freshwater interface position (based on available chemical data) to calibrate the model and verify the hydraulic parameters.

Thesis Outline

This thesis is organized into six chapters. Chapter 1 describes the main reason for conducting the present study, as well as introduces the study region area, describes previous studies conducted in the region, and summarizes background literature on this subject.

Chapter 2 describes the hydraulic parameters which are required as input to numerical model. Chapter 2 also describes geology of the area, the methodology used to determine the hydraulic parameters, and provides a range of estimates that are used as input to the models. Chapter 3 includes background on saltwater intrusion, a brief description of the code used for numerical modeling, and the conceptual model. Results of simulations for flow and solute transport, which correspond to current actual position of the saltwater–freshwater interface, are included in Chapter 4.
Chapter 5 integrates all data and includes a summary of the history of glaciation during Pleistocene in southwest BC and on Saturna Island. The chapter also provides the results of transient simulations that aim to determine if the theory of saltwater movement for the last 11,000 years is consistent with the sea level history. Chapter 6 provides conclusions and recommendations for future work. The appendices include aerial photographs with descriptions of the hydraulic test data and the effect of fracturing on the results.
CHAPTER 2
HYDRAULIC PARAMETERS - HETEROGENEITY
CHARACTERIZATION

Previous studies involving saltwater intrusion modeling (e.g., Huyakorn et al., 1987; Voss and Souza, 1987; Pinder and Cooper, 1970) demonstrated the importance of having a good knowledge of hydraulic parameters. Heterogeneity, or the spatial variation in hydraulic parameters, controls the pattern of groundwater flow, and is especially important when there is associated transport of solutes and/or density dependent flow (Dagan, 1989). Most published modeling studies have been conducted for isotropic and homogeneous media, but a few studies (e.g., Dagan, 1989 and Simmons et al., 2001) investigated the role of heterogeneity on variable density flow. Those studies used a stochastic approach to model aquifer heterogeneity in porous media.

This chapter provides an overview of fracture flow and the various methods that can be used to model groundwater flow in heterogeneous fractured media. The concept of hydrostructural domains (as described by Mackie, 2002) is described within the context of modeling groundwater flow in the study area. The chapter also summarizes work that was carried out as part of this research to quantify the hydraulic properties for the Gulf Islands geologic formations.

Fracture Flow Models

Fractured aquifers are more complex and difficult to model than porous media because the flow is controlled by fractures that represent multiple and discrete paths. A fracture is a surface of discontinuity, and includes cracks, joints and faults. Mechanisms that produce fractures and factors that determine their characteristics have been studied for a long period of time in different geological settings by various researchers (e.g., Suppe, 1985; Pollard and Aydin, 1984). Brittle failure deformation produces three types of fractures (Twiss and Moores, 1992):

- Extension fractures form perpendicular to the minimum stress and parallel to the maximum stress,
Tension fractures (extensional) are produced in response to a minimum tensile stress,

Shear fractures - formed under triaxial compression.

Both the evolution of the stress field and the fracture network control groundwater flow as the flow is mainly concentrated within open and connected fractures. Flow in fractured aquifers is further controlled by the characteristics of fractures: strike and dip, density, aperture, roughness, and degree of connectivity. The interconnectivity between fractures can provide paths for groundwater flow and increases the permeability of a rock mass, or it can reduce primary porosity by sealing faults (Suppe, 1985).

Fractured aquifers are characterized by fractures, which typically have a high permeability, and a rock matrix, which typically has a lower permeability relative to the fractures. When the rock matrix has a low permeability, flow occurs within the fractures, while groundwater storage is within the matrix. Flow between fractures and the matrix is determined by the permeability gradient, which is related to fracture aperture, fracture wall roughness, and channelling effects that induce phenomena like dispersion, diffusion, or sorption effects at the fracture–matrix interface. Generally, the presence of fractures induces a secondary porosity to a medium.

Various researchers have studied fractured aquifers from different perspectives. Bear and Berkowitz (1987) summarized these approaches by defining four scales for flow in fractured aquifers:

- Very near field – flow occurs in a single fracture,
- Near field – flow occurs in fractures and each fracture is described in detail.
- Far field – flow occurs in two overlapping media.
- Very far field – flow occurs in an equivalent porous medium.

Therefore, describing flow in fractured aquifer can be approached using one of three basic approaches: 1) discrete flow or flow within a single or network of individual fractures, 2) dual (or double) porosity flow within the fractures and the matrix, and 3) equivalent porous media involving the calculation of equivalent hydraulic parameters for the entire domain representing a continuum of factors.
When analyzing hydraulic test data to determine the hydraulic parameters of an aquifer, or when developing a flow model, it is important to determine a hydrogeologic model that best describes the geology, structure (i.e., fractures) and flow patterns. An important factor in choosing a suitable model is the scale of study, specifically, microscale or macroscale, where the latter is typically based on the concept of the representative elementary volume. At a small scale, where it is of interest to obtain high-resolution results (such as would be required in contaminant transport modeling), it is necessary to have a large amount of data and a high degree of spatial resolution on fracturing. This requires detailed measurements of all fractures and an accurate representation of the spatial distribution of fractures over the entire area. In contrast, large-scale studies involving regional groundwater flow require less spatial resolution, because local variations would only serve to complicate the general flow behaviour. In this case, the equivalent porous media approach to modeling fractures has been more successful.

Both numerical and analytical models can provide solutions for aquifer behaviour under a variety of hydraulic stress conditions, and consequently, both are used to estimate the hydraulic parameters of aquifers (inverse modeling) and to simulate flow (forward modeling). Numerical models for fractured media involve a large amount of data and a description of individual fractures; a huge computational effort. The results are more precise, and give a good representation of the aquifer as a whole, but they become inaccurate because of the many uncertainties in the model. Analytical models similarly provide mathematical solutions, but are less flexible than numerical models because analytical solutions are difficult or impossible to obtain for situations arising from complex boundary conditions or non-uniform distribution of hydraulic properties (e.g., heterogeneity). Nevertheless, there are some solutions for simple fractured aquifer geometries (e.g., Barenblatt et al., 1960; Warren and Root, 1963). In cases where there are a large number of data that describe aquifer heterogeneity, stochastic methods can be used to characterize the entire region.
Discrete Fractures

Analytical models have been developed to describe flow in the vicinity of a single discrete fracture (e.g., Gringarten, 1982) or in uniformly fractured media (Bourdet and Gringarten, 1980; Kazemi et al, 1969; Warren and Root, 1963). Numerically-based discrete fracture models were derived using principles of rock mechanics and are designed to describe the geometry of a fracture network and fracture generation. Baecher et al. (1977) and Dershowitz (1992) developed conceptual models to describe fractured media based on fracture generation. That approach is suitable for small-scale aquifers, and describes, in detail, the flow through fractures and between connected fractures.

Dual Continuum Media

Barenblatt et al. (1960) and Warren and Root (1963) described the flow in fractured aquifers resulting from two overlapping medium: a porous medium consisting of lithified sediments, and a superimposed network of fractures. Figure 2.1 shows a conceptualization of this model, where each medium is characterized by different properties. The porous media (solid rock matrix) has a high storage capacity and low permeability, while the fractures have a low storage capacity, but high permeability. The water is stored mostly in the pores of the rock, while flow occurs mainly through fractures, as well as across the interface between the fracture and porous medium, or between connected fractures. Streltsova (1978) described the flow between fractures and porous medium in detail. Transfer of fluid between fractures and pores is a slow process, and although it is generally assumed to be at steady-state, this is unlikely. In the case of a dual continuum media, two different hydraulic conductivities are assigned - one for fractures and one for rock.

Equivalent Porous Media (EMP)

When numerous small, interconnected fractures are present over a large scale area, the most widely-used model for the aquifer is based on the equivalent porous media approach. In this case, the fractured aquifer is treated as a continuum of fractures and rock matrix, and a single hydraulic conductivity is assigned to the entire aquifer, as
Figure 2.1 Equivalent porous media versus double porosity representations of fractured bedrock (dual porous fractures and matrix are characterized by different values of hydraulic conductivity, and the flow occurs mainly through fractures and also between the two media, equivalent porous concept - a single hydraulic conductivity value is assigned for both matrix and fractures).
shown in Figure 2.1. The approach is based on the being able to identify an appropriate representative elementary volume (REV) for which equivalent hydraulic properties can be defined. In practice, it is difficult to determine the REV, but often the results from pumping test analyses are taken to represent a volume of aquifer of appropriate size to reflect the REV. Analytical models yield a single value for the hydraulic parameters for both rock matrix and fractures. The water is assumed to flow through pores as well as through fractures, and a large volume is characterized by the sum of all fractures and porous media permeabilities. Therefore, approximations used in describing fractured rocks depend on the scale of investigation, and require definition of a representative elementary volume (REV).

Guerin and Billaux (1994) concluded that the REV for transport processes can be different from the REV for flow. It has been suggested that the REV model is a poor predictor of spatial distribution of fluxes through fractures and is a poor predictor in the case of transport flow. Thus, for the case of solute transport in a coastal aquifer, the magnitude of saltwater intrusion may be sensitive to heterogeneities. At a small scale, the pattern of fracturing will determine paths for groundwater flow and for saltwater to intrude the aquifer. Consequently, the choice of an EPM approach to model density-dependent flow in a coastal aquifer may not adequately reproduce transport at a small-scale, and may not accurately represent the flow and transport processes within the fracture network. However, at a large scale the heterogeneity introduced by fracturing may be less significant and overall trends may be adequately produced.

Previous studies indicate successful results by considering EPM to describe heterogeneities in karst systems (Scanlon et al., 2003), layered aquifers (Kuniansky, 1993), and fractured rocks (Dougherty, 1985). Pruess et al. (1986) showed that the EPM model is not applicable in the case of large fracture spacing, nor in cases where the matrix has a low permeability. Both of these conditions exist on the Gulf Islands as fracture spacing in some areas can be quite large, and the matrix is of low permeability. Consequently, the EPM approach may be further limited in its application in this study. However, as noted earlier, if the scale of the study is sufficiently large, an EPM approach may be justified.

Bai et al. (1993) introduced the concept of multi-porosity / multi-permeability by considering the fact that the entire domain may be characterized by zones with different hydraulic properties. In this case, multiple distinct sub-systems are assigned in the
common domain. In many respects, this approach is similar to the hydrostructural domain approach proposed by Mackie (2002) on the basis of structural mapping (discussed in the next section).

**Fracture Models for the Gulf Islands**

**Fracture Mapping Study**

In the Gulf Islands region, water wells are drilled in fractured rocks, and past investigations indicate that primary porosity is low and that flow is mainly through fractures (Dakin, 1983). The Gulf Islands region is dominated by northwest trending ridges that are the result of deformation, uplift and erosion since Late Jurassic to Holocene (Journeay and Morrison, 1999). Previous studies (Journeay and Morrison, 1999; England, 1990) identified north trending structures that cross-cut fold axes in Gulf Islands. Structurally, the area is characterized by 1 to 5 metre fracture spacing with a displacement of less than 10 meters, and northwest trending faults with an offset of 500 meters. At regional scale, major structural features have a NE-SW orientation.

Mackie (2002) used a scan-line technique (i.e., measuring the fractures along a measuring tape) to map fractures on the Gulf Islands. Mackie (2001) identified three key types of fractures on the Gulf Islands:

1. Bedding perpendicular joints – with mostly NE-SW maximum compressive stress orientations. This type of fracture is the most common and includes both mineralized and non-mineralized joints. Bedding perpendicular joints are more closely spaced in the mudstone-dominated formations and spacing increases in sandstone dominant formations; the variation in joint spacing is attributed to bedding thickness (Suppe, 1985).

2. Discrete fractures and faults – generally vertical to sub-vertical and numerous throughout the Gulf Islands. A major fault is present on Saturna Island (the Harris Fault), which trends north-northeast at a dip of 45°. One structural characteristic associated with the presence of a fault is that fracture density increases near the fault plane, particularly in the sandstone-dominant formations.
3. Low angle fractures – assumed to be related to isostatic rebound following Quaternary deglaciation, or Neogene uplift.

Most of these fractures are heterogeneously distributed with respect to both lithology and geological structures, and fracture density varies widely for entire Gulf Islands (Mackie, 2002). Of these three fractures types, discrete fractures (bedding perpendicular joints and low angle fractures) are believed to represent conduits for groundwater flow rather than barriers to flow, and several orientations should be open to flow in consideration of the current tectonic stress regime. The density of these discrete fractures, determined on the basis of their spacing, is not constant, but it is related to proximity to major faults. Low angle, bedding plane fractures are expected to be open to flow as a result of isostatic uplift following deglaciation.

Understanding and quantifying fractures on the Gulf Islands implies knowing their location, density, direction, degree of connection, and properties for conducting flow. The main questions to be answered are: What is the relationship between hydraulic testing response and presence of fractures? Which set of fractures allows groundwater flow? The answer to the first question may be found by considering the hydraulic responses to pumping; this will be addressed as part of this study. The answer to the second question involves consideration of flow in three-dimensions with specific emphasis on fracture orientations and the existing stress regime (a topic beyond the scope of this study).

**Modeling Approaches**

Because the primary porosity of the rocks is very low (<5%; England, 1990), the dual continuum model is not expected to be valid on the Gulf Islands. Therefore, there are two potential approaches for determining the hydraulic parameters and undertaking flow and transport modeling. These are the discrete fracture approach and the EPM approach. For hydraulic property determination, the following methodologies are possible:

1. Knowing the exact location of all fractures intersecting the borehole that is tested hydraulically, and undertaking packer testing to determine the permeability of each fracture – Discrete Fracture Approach.
2. Using the measurements of fracture characteristics to estimate an equivalent hydraulic conductivity ellipsoid for the aquifer (e.g., using FracMan) – Discrete Fracture/ EPM Approach.

3. Using well test data to estimate equivalent porous media properties for the aquifer – EPM approach.

Given the size of the area, a discrete fracture approach based on the measurement of the properties of individual fractures is unrealistic. The discrete fracture approach based on determining equivalent properties for the aquifer is a realistic approach, but will not be undertaken as part of this study.

For flow and transport modeling, the choice of code for this thesis limits the range of possible methods. Specifically, USGS Sutra will be used, and this code is an EPM-based code. Therefore, equivalent properties must be determined for the aquifer and input into the model for simulating flow and transport. Multiple open hole aquifer tests were analyzed to estimate the hydraulic parameters of the aquifer using several analytical EPM-based methods (discussed later in this chapter). For flow modeling, these parameters are assumed to represent the continuum of fractures and rock matrix, consistent with the EPM concept. The EPM approach is considered appropriate for several reasons: a relatively large scale of investigation, a small amount of discrete data, and a long simulation time. However, because of the complexity of fracturing observed on the Gulf Islands, as described in the following section, it is necessary to try and capture the variability observed from formation to formation and from one area to another. To do this, a hydrostructural domain approach is used.

**Hydrostructural Domains**

Previous studies of groundwater flow in the fractured rocks of the Gulf Islands identified a need to represent flow in the fractured aquifer and the variability in flow caused by spatial variations in fracturing. Abbey (2001) identified two types of hydrostratigraphic domains: unconsolidated overburden and bedrock, based on geophysical surveys and hydraulic testing. Following a detailed fracture mapping study, Mackie (2002) expanded that conceptual model to incorporate the concept of
hydrostructural domains. Mackie described a hydrostructural domain as: “an area or volume in which fracture intensity or density is significantly different from areas or volumes outside this space, and in which groundwater parameters, such as storage or transmissivity, may be different from the area or volume outside as a direct result of the difference in fracture intensity”. Figure 2.2 (from Mackie, 2002) captures the key elements of the hydrostructural domain conceptual model. The model includes a lithologic-based domain and structure-based domain, which are defined as follows:

**Lithology-based hydrostructural domain**: consists of mudstone- and sandstone-dominant formations. Classically, mudstones have a lower permeability than sandstones, but a high density of bedding perpendicular joints in thinly bedded mudstone units on the Gulf Islands results in an increase in permeability. Stratigraphic relationships result in three bedding thicknesses: thick beds (sandstone), intermediate (mudstone) and thin (transition zone between mudstone to sandstone), and consequently, three degrees of fracturing. Therefore massive sandstone formations have lower fracture intensity than the mudstone formations, and the inter-layers are highly fractured due to multiple layering.

**Structure-based hydrostructural domain**: is a consequence of the density and aperture of fractures. It consists of discrete faults and fractures that are widely distributed and major faults and fracture zones where the fracture intensity increases in the proximity of the major fault or fracture zone.

In this research, cross-sectional models will be created to model solute transport. Therefore, the three-dimensional character of the aquifer cannot be captured. This means that only lithologic variations, as represented by the lithology-based hydrostructural domains, can be modelled. Therefore, lithologic units (massive sandstone, massive mudstone and interbedded sandstone and mudstone) will be characterized by equivalent hydraulic properties in the range of calculated values based on the degree of fracturing present (a continuum approach). The properties for each should be estimated by considering a large scale of investigation (pumping tests) and major characteristics for each formation.

In three dimensions, the locations of discrete fractures, faults and fracture zones could be modeled, but this is beyond the scope of this research. This would involve representing the structure-based hydrostructural domains.
Hydrostructural Domain,  
(after Mackie, 2002)

<table>
<thead>
<tr>
<th>Fault and Fracture Domain</th>
<th>Discrete Fault and Fracture</th>
</tr>
</thead>
</table>

<table>
<thead>
<tr>
<th>Hydrostratigraphic units (after Abbey, 2001)</th>
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</thead>
<tbody>
<tr>
<td>Overburden</td>
</tr>
<tr>
<td>mudstone-dominant lithology</td>
</tr>
<tr>
<td>Bedrock</td>
</tr>
<tr>
<td>sardstone-dominant lithology</td>
</tr>
<tr>
<td>Increasing relative permeability</td>
</tr>
</tbody>
</table>

Figure 2.2. Hydrostructural domain-conceptual model (modified from Mackie, 2002).
At a small scale, discrete fractures would be more transmissive than the surrounding rock matrix and would best be represented as discrete fractures because of their wide spacing. At a larger scale it would be possible to assign equivalent hydraulic parameter values depending on the proximity of major fault and fractures or based on the type of lithology. Previous studies on Saturna Island (Mackie, 2001) indicated that the porous medium can be separated into a few domains characterized by different hydraulic properties. One domain would be defined by high permeability, with permeability increasing as the density of fractures increases near a major fault or fracture zone. The rest of the region would have a lower permeability, reflecting a continuum between the primary rock permeability and the secondary fracture permeability. The lower hydraulic capacity in the region surrounding the fault would be due to the greater spacing between fractures, and lower degree of connectivity.

**Hydraulic Testing - Determining Hydraulic Parameters**

Hydraulic parameters typically are derived by hydraulically testing an aquifer and analyzing the data. Hydraulic data can be obtained from aquifer tests (including constant discharge tests and recovery tests), slug/bail tests or tidal tests. Aquifer tests consist of applying a stress to an aquifer by pumping a well, and monitoring the response of the aquifer in that well and any observation wells. The response is measured by the amount of drawdown in the well as a function of time. The test data are then compared to the type of response predicted by an appropriate analytical model, and the hydraulic parameters, transmissivity (T) and storativity (S), are derived. Transmissivity represents the rate of flow that moves in unit time, under a unit hydraulic gradient through a saturated unit area of the aquifer. Hydraulic conductivity (K) can be determined directly from T as it represents the transmissivity per unit saturated thickness of the aquifer (Kruseman and deRidder, 1994). Storativity (S) represents the amount of water that is released from storage from an aquifer per unit area per unit decline in hydraulic head.

Analytical models generally assume an ideal aquifer – radial flow, confined, homogeneous and isotropic aquifer, with large aerial extent. As these particular conditions do not apply to the Gulf Islands, it is necessary to determine a realistic conceptual model that accurately represents the type of flow observed during the test,
and then to calculate hydraulic parameters, T and S, which are representative of the fractured aquifer and can be input into a numerical model.

Previous geologic and hydrogeologic studies on the Gulf Islands showed a high degree of variability of geologic and hydraulic features both in the different formations and within the same formation. Allen (2001) and Mackie (2002) pointed out that higher hydraulic parameters are associated with presence of faults and fractures. Abbey (2001) concluded that most of the results are scale dependent, with hydraulic parameters depending on the length of the test and the location of the well relative to faults. T and S values were observed to decrease with the duration of aquifer test, and increase with fault proximity.

Even though a number of analytical methods for different types of aquifers were previously used, none are entirely appropriate for heterogeneous aquifers. Each fractured formation has its own particular characteristics with regard to fracturing, which differs from one area to another. Therefore, no single model can be uniformly applied to all formations. Thus, both careful examination of the character of the flow regime and the application of multiple methods for determining hydraulic parameters are the approaches used in this study. In this way, the most appropriate model can be selected based on the geologic environment particular to the aquifer test, and a comparison of the parameter estimates obtained from each model can be made.

The following tasks were undertaken:

1. To identify appropriate conceptual models for the Gulf Islands that represent the heterogeneities anticipated in the region.
2. To determine hydraulic parameters using data from aquifer tests. The hydraulic parameters, T and S, can be determined quantitatively by applying analytical models in an inverse modeling fashion, or qualitatively, using derivative methods that allows for the identification of the type of flow regime: radial or linear. The flow regime provides insight about presence of fractures.
3. To compare the results obtained by different analytical methods to determine a possible correlation between the methods. Results for log T and log S will be plotted to determine the trends and the accuracy of the various methods.
4. To locate the wells on air photos and map geologic and structural features. This will allow determination of the relationship between the wells and the structural features (faults and fractures).

5. To complete a database containing estimates of the hydraulic parameters obtained by analyzing long- and short-duration pumping tests.

6. To assign proper hydraulic parameters to the Nanaimo Group formations. Most of the tests include a well log and a map indicating the location of the well, based on which the wells can be related to a geologic formation.

The Database

The data for this research are from constant discharge tests and recovery tests, and were provided either by previous research studies, by studies conducted as part of this research, or by studies conducted by consultants and are presented as an enclosed CD in Appendix A. Most of the tests were conducted on domestic wells to determine the long term well yield (maximum pumping rate).

All data in this thesis were compiled, analyzed and interpreted as part of a study to assess the hydraulic parameters for the entire Gulf Islands region (Allen et al., 2003). The analysis of the data for that project formed a major component of the scope of work for this thesis, and therefore, the results are summarized herein.

Six aquifer tests were collected, analyzed and synthesized from previous reports (Allen, 2001 and Abbey, 2001). The qualitative characteristics of flow regime were determined by identifying different types of flow and boundaries using geophysical and hydrogeological methods, T and S values were calculated. Aquifer test data were largely provided by the BC Ministry of Water, Land and Air Protection and by consulting companies. These included 11 long duration tests and 20 short duration tests.

Many aquifer tests provided data for both the pumping well and one or more observation wells, and data were collected during pumping and recovery. Several tests were conducted for pumping wells only. The tests are distributed among the geologic formations as indicated in Table 2.1. The degree of confidence in the results is higher for larger sample populations. Not all formations on the Gulf Islands or on Saturna Island are represented due to lack of available testing wells.
Table 2.1 Number of aquifer tests by geologic formation

<table>
<thead>
<tr>
<th>Formation</th>
<th>Gabriola</th>
<th>Geoffrey</th>
<th>Spray</th>
<th>de Courcy</th>
<th>Cedar District</th>
<th>Protection</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nr. of tests - long duration</td>
<td>12</td>
<td>13</td>
<td>3</td>
<td>4</td>
<td>2</td>
<td>3</td>
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<tr>
<td>Nr. of tests - short duration</td>
<td>61</td>
<td>9</td>
<td>5</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
</tr>
</tbody>
</table>

When conducting a study to determine hydraulic parameters or hydrogeological boundaries (e.g., faults) one should follow a pre-established testing program or should have a good distribution of the tests (Kruseman and de Ridder, 1990). The tests conducted on the Gulf Islands are randomly distributed and have different lengths.

The tests were separated into two groups based on the length of pumping test, as previous studies indicated a sensitivity of the well responses and hydraulic parameters to length of test. Tests longer than few hours were considered as long-duration tests, and tests conducted for few minutes to one hour were considered short duration tests.

A previously-developed Excel database (Allen, 1999) was used in this study to maintain consistency with previous work. Single well and multiple wells tests were organized into separate computer directories (one per well site). A total of 31 directories contain the test data for the entire region. Data provided in Appendix A are organized in directories that contains subdirectories of multiple pumping and observation wells at each site, and the pumping and recovery test data for each well, if available. The main spreadsheet page (Figure 2.3) contains information about well construction, initial water level, discharge rates, and water level measurement made at different times. The other pages in the spreadsheet contain plots and associated methods of analysis, and the results. The following lists the various spreadsheet file's contents:

1. Drawdown and time data.
2. A log-log graph of drawdown versus time (for use with the Theis method and for identifying types of flow).
3. A semi-log plot of drawdown versus log-time (for use with the Cooper-Jacob methods 1 and 2).

4. The derivative calculation.

5. A log-log plot of the first derivative of drawdown versus time (to identify the radial flow period and wellbore storage time interval).

6. A summary of T and S values obtained using each method.

Methodology

According to Kruseman and de Ridder (1999), conventional porous media analytical methods can be used in the case of low-permeability bedrock having a high density and uniformly-distributed fractures. In this case flow occurs only through fractures, but the aquifer behaves as a consolidated homogeneous aquifer. However, there are certain limitations to applying porous media models (as described below). The literature also provides a few models that were developed specifically to consider flow in fractured rocks. As discussed previously, there are two conceptual models: the double porosity (dual continuum) model discrete fracture models (including the single vertical fracture model).

Qualitatively the response of an aquifer to pumping at a constant discharge rate can be described in terms of the type of flow regime. If the aquifer is homogeneous and isotropic, of uniform thickness and of infinite extent, then the response is radial, and a conventional method of analysis, such as Theis or Cooper-Jacob can be used. When there is a hydraulic discontinuity, such as a single fracture that passes close to a pumped well, linear flow is commonly observed at early times (Figure 2.4.). During this time, flow comes mainly from storage in the fracture. On a log-log plot of drawdown versus time, the slope of the line is approximately 0.5, which accounts for the linear flow associated with the single fracture (Gringarten, 1982). At later times (mid to late time) radial flow can occur, and the flow to the well is supplied by the entire aquifer, not only by the fracture.
**FILENAME: ptd Pumping ell (001; Well 9).xls**

**AQUIFER TEST DATA**

**Well: WELL 9**  
Date of test: OCT. 23 - NOV. 3, 1970

**Type of aquifer test:** CONSTANT DISCH. (Q)

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<th>How Q measured</th>
<th>WATER METER (w/ sweep hands)</th>
<th>Well type:</th>
<th>PUMPING</th>
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</table>

Discharge rate (Q): 20 IGPM; 25 IGPM AT 650 MIN.

Meas. point for w.l.'s:  
Top of Casing  
Length of test:  9517 MINUTES

Stick-up of casing: 0.5 FEET (infered)

Static water level: 171.9 FEET ASL (16.3 feet below top of well casing)

Well Depth:  220 feet

Note: Original data in feet.

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<th>Drawdown (feet)</th>
<th>Water Level Reading (meters)</th>
<th>Drawdown (meters)</th>
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</tbody>
</table>

Figure 2.3 Spreadsheet including information about wells construction, discharge rates, water levels.
If the hydraulic parameters of the aquifer are to be determined from data that are influenced by a fracture, then it is necessary to use an appropriate linear flow model (e.g., Gringarten, 1982) or apply a conventional radial flow model to the appropriate radial flow data (Allen, 1996). A proper method to identify the radial flow period is the derivative method (Spane and Wurstner, 1993). This method uses a graph of the first derivative of drawdown versus log time; the radial flow period corresponds to a horizontal line as illustrated in Figure 2.5. Radial flow typically occurs at the mid-portion of the drawdown curve, and reflects flow coming from both the fracture and the surrounding aquifer. The flow is described as pseudo-radial. If the fracture is hydraulically significant linear flow may persist for a longer period and pseudo-radial flow may not be observed if the test is too short.

Many of the tested wells were drilled by consultants, and therefore, many are in close proximity to major faults or fracture zones which yield large quantities of groundwater. For this reason, it can be expected that many tests may exhibit linear flow. In fact, Allen (1999) found that most long duration pumping tests conducted in fractured bedrock in BC exhibit linear flow. Because of the heterogeneous geology and structure of the area and the random distribution of the tests, no single analytical model is expected to be applicable for all cases. Thus, using multiple methods of analysis, which represent the range of flow regimes anticipated for the region, may provide a means to examine the types of flow at each site. Considering the range of fracturing found in the Gulf Islands and the range of pumping responses associated with each type, the following analytical methods were used:

1. Derivative method – to identify different types of flow regime, especially radial flow. This is done to be able to apply the Theis and Cooper-Jacob methods.
2. Theis method – a confined, homogeneous aquifer model used only during the radial flow period to calculate $T$ and $S$ values.
3. Jacob-Cooper method - a confined, homogeneous aquifer model used only during the radial flow period to calculate $T$ and $S$ values.
5. Gringarten-Whiterspoon (1972) – single vertical fracture model for observation wells near a pumped well drilled into a vertical fracture.

6. Theis recovery method – a confined, homogeneous aquifer model used during the radial flow period.

Each method is described in more detail in the following subsections.

**Derivative Method**

Characterization of heterogeneities is the first step to understanding and predicting the behaviour of groundwater flow to pumping. Bourdet et al. (1984) introduced the concept of derivative curves a tool that would allow for identifying types of flow regime, as well as hydrogeologic boundaries, leakage and wellbore storage (Spane and Wurstner, 1993). This theory is based on the fact that the pressure derivative is sensitive to small variations of pressure.

The derivative method was the first step used for analyzing aquifer test data in this study. It was used to identify the period of flow associated with wellbore storage effects, and to differentiate linear flow from radial flow. Wellbore storage is associated with a release of water from the well itself, not the aquifer. Consequently, it is desirable not to consider this period of flow.

Figure 2.5 shows an example of the type of response that usually appears in bedrock wells tested in the Gulf Islands. Wellbore storage occurs mainly in pumping wells and dominates the first part of data. Wellbore storage is manifest as a “hump” on the derivative graph or as a straight line with a slope of one on the log-log drawdown versus time plot (Figure 2.4). Following wellbore storage is a period of linear flow, which evolves to radial flow (characterized by a flattening of the derivative curve). At late times, many tests indicate the presence of a boundary. Boundaries act either to increase (as in the case of a recharge boundary) or decrease (as in the case of a barrier) the rate of change of drawdown measured in the well. If the aquifer is infinite-acting (of large unobstructed extent) the derivative graph maintains a horizontal appearance.

Radial flow may be identified on log-log plots of drawdown versus time, but the presence of linear flow, changes in pumping rate, or leakage, complicate the curve.
Figure 2.4 Linear flow occurs at early time, followed by radial flow marked by afflattening of the curve.

Figure 2.5. Log-log plot of the derivative of drawdown versus time and corresponding types of flow.
Therefore the derivative method is a valuable method to differentiate each type of flow before applying an analytical model.

**Theis Method**

Theis (1935) developed the first analytical model for unsteady-state radial flow in confined aquifers, and the method consists of curve matching. The method requires a constant pumping rate, a confined aquifer of infinite aerial extent, homogeneous and isotropic hydraulic properties, and a constant aquifer thickness. In addition, the well must fully penetrate the aquifer, and the initial potentiometric surface must be horizontal (Kruseman and de Ridder, 1999).

The Theis method consists of plotting on a log-log scale the drawdown versus time for the pumping well and any observation wells. For curve matching, a second graph with identical scale (the Theis curve for 1/u versus W(u)), is plotted. Maintaining the axes parallel, the Theis curve and the drawdown curve are hand matched. An arbitrary match point is defined using the following coordinates: time (t), drawdown (s), W(u) and u. The dimensionless parameters u and W(u) are defined in equations 2.1 and 2.2.

\[ u = \frac{Sr^2}{4Tt} \]  
\[ W(u) = 0.5772 - u + u - \frac{u^2}{2 \cdot 2!} + \frac{u^3}{3 \cdot 3!} - \frac{u^4}{4 \cdot 4!} + \ldots \] 

Equation 2.3 and 2.4 are used to solve for T and S.

\[ T = \frac{Q}{4\pi S} W(u) \]  
\[ S = \frac{4uTt}{r^2} \]

Where:

T= transmissivity (m²/s)

S= storativity (dimensionless)
Q= pumping rate (m³/s)

s= drawdown (m)

t= time (s)

r= distance between observation well and pumping well, or effective radius for pumping well (m)

Although technically not a fractured aquifer flow model, Theis can be used to calculate T and S. Allen and Michel (1998) demonstrated that if only the radial flow period is used, the Theis method provides reasonable estimates of the aquifer parameters when the flow regime is influenced by linear flow associated with faults or fracture zones. Therefore, the derivative method was used first to identify the radial flow period, and the appropriate time interval for the data were analyzed using the Theis method.

Figure 2.6 shows a graph of drawdown versus time for a pumped well. At times less than 1 minute, wellbore storage occurs; this time period was not included in the analysis. Linear flow occurs from 1 minute to about 10 minutes as identified on the derivative graph. The slope of the curve is 0.5 for this linear flow period. Radial flow occurs from 10 minutes to 100 minutes. Unconfined behaviour is observed in this well from about 100 minutes until the end of the test (flattening and then rise in drawdown).

In the observation wells, linear flow dominates the early part of the tests and is followed by radial flow. In some cases, the presence of a boundary is marked by an increase in drawdown. Radial flow in observation wells is sometimes delayed, and the flow is dominated by the linear flow, which indicates a high connectivity between observation and pumping well. For example Figure 2.7 shows the response of an observation well connected to the pumping well by a major fracture. The response is mostly linear, and indicates that radial flow does not occur, even at late time. In other cases, when the well is pumped for a sufficient period of time, radial flow occurs over the same time interval in both the pumping and observation wells.

Regardless the fact that most wells are influenced by linear flow, the Theis method yields similar estimates of T and S compared to the other methods, including the fracture-based models, if only the radial flow period data are used for analysis.
Figure 2.6 Drawdown variation with time, pumping well, Theis method.

Figure 2.7 Drawdown variation with time, observation well, Theis method.
Jacob-Cooper Method

Under the same conditions outlined previously for Theis, but for small values of \( u \), plots of drawdown versus time on a semi-log plot will yield a straight line. The Jacob-Cooper (or equivalently, the Cooper-Jacob, or Jacob) Method is a simplification of Theis, and because it does not require curve-matching, is the most frequently-used method for pumping test analysis. The Jacob-Cooper method (1964) was derived from the Theis method, and it is applicable for an unsteady-state flow, for values of \( u < 0.01 \) (i.e., for small radial distances between pumping and observation wells or for sufficiently large pumping times). Linear regression allows for determining the slope and x-intercept (time intercept) of the straight line. These values are used to calculate \( T \) and \( S \) using:

\[
T = \frac{2.3Q}{4\pi \Delta s} \quad (2.5)
\]

\[
S = \frac{2.25T t_0}{r^2} \quad (2.6)
\]

Where:

- \( T \) = transmissivity (m\(^2\)/s)
- \( S \) = storativity (dimensionless)
- \( Q \) = discharge rate (m\(^3\)/s)
- \( \Delta s \) = slope of the drawdown curve on a semi-log scale (or drawdown difference over one log cycle) (m)
- \( t_0 \) = x-intercept of the drawdown curve on a semi-log scale (s)
- \( r \) = radius of the well (m)

Semi-log plots, as shown in Figure 2.8, display the same type of responses as previously indicated for Theis. Two applications of the Jacob-Cooper method were used:

1. Only the radial flow portion of the curve (same as Theis method) was used to calculate the slope and intercept (method 1),
2. The entire part of the curve was used to calculate slope and intercept (disregards the presence of linear flow in the data) (method 2). In both cases, the first part of the graph was discarded if the effects of wellbore storage were present.

To determine the effect of bias introduced by analyzing different portions of the drawdown, the two sets of results were compared. A plot of the log T values obtained using Jacob-Cooper method 1 versus method 2 (Figure 2.9) revealed a close agreement, as shown by the regression equation $y = 1.0072x + 0.2221$, and $R^2$ value which is close to unity. T values were slightly higher when the entire set of data was considered. Because the entire data set includes the linear flow period, the T values for the aquifer would be biased and reflect the presence of fractures.

**Theis Recovery Method**

Recovery data are an important source of information from aquifer tests. These often provide more accurate data because the data are not affected by changes in pumping rate. Recovery testing consists of recording the recovery of the water level after shutting off the pump. The water level recovery is referred to as residual drawdown, and is calculated by subtracting the recovery drawdown from the initial water level. Residual drawdown is plotted against the ratio of the total time to the recovery time ($t / t'$) as shown in Figure 2.10. Theis recovery method is based on the Principle of Superposition, which assumes that after shutting off the pump, the well continues to be stressed by an injection well superimposed on the pumping well, with an injection rate equal to the previous discharge rate, according to equation 2.7

$$s' = \frac{Q}{4\pi T} \left(W(u) - W(u')\right) \quad (2.7)$$

Where: $u = \frac{r^2 S}{4Tt}$ and $u' = \frac{r^2 S'}{4Tt'}$

and

$r =$ radial distance from the pumping well (m).

S and S' are the storativities during pumping and recovery, respectively, dimensionless.
Figure 2.8 Semilog plot of drawdown versus time (Cooper - Jacob method).

Figure 2.9 Comparison between T values obtained from Cooper - Jacob method 1 versus method 2.
\[ T = \text{aquifer transmissivity (m}^2/\text{s)}, \]

\[ t \text{ and } t' = \text{total time and recovery time after pumping.} \]

For small values of \( u \) and \( u' \), \( T \) is constant, and the storativities are equal. As a consequence:

\[ s' = \frac{2.3Q}{4\pi T} \log \frac{t}{t'} \quad (2.8) \]

\[ T = \frac{2.3Q}{4\pi \text{slope}} \quad (2.9) \]

In a homogeneous, isotropic, confined aquifer that is unbounded, a plot of \( s' \) versus \( t/t' \) on a semi-log scale will yield a straight line, for large values of time. The slope of this line is used to obtain a value for \( T \), according to equation 2.9. Theis recovery analysis does not provide a value for storativity.

As neither the plot of drawdown versus time nor the derivative graphs exhibited an extensive radial flow period, the recovery method is not entirely applicable. As for the previous analyses, the same methodology was applied in this case. Only the data that correspond to the radial flow period or equivalently, the same time period for the pumping test, were analyzed.

Transmissivity values obtained from recovery data are slightly lower than the ones obtained by applying the Theis method (Figure 2.11). However recovery method is not influenced by the pumping rate and is a response of the well to flow through fractures and through pores.

**Single Vertical Fracture – Gringarten-Witherspoon Method**

A fracture that is superimposed over a porous media modifies the flow pattern and hydraulic characteristics of the aquifer. The single vertical fracture model is based on the following assumptions: a homogeneous, isotropic, infinite extent aquifer is crossed by a single vertical fracture over its entire thickness; the aquifer is confined by two horizontal impermeable beds. The hydraulic conductivity of the fracture, with zero width and no storage, is assumed infinite, compared to the conductivity of the porous media. Therefore, the drawdown will be uniform for the entire length of the fracture.
Figure 2.10 Semilog plot showing the recovery behavior influenced by the fractures presence at early times and radial flow later on.

Figure 2.11 Comparison between hydraulic properties determined by using pumping data (Theis method) or recovery data (Recovery method).
At early times, the flow is one-dimensional, and linear, entering perpendicular on the fracture along its entire length. The drawdown in this case is a function of perpendicular distance from the fracture. As the pumping influence advances away from the well, the flow becomes pseudo-radial and water comes from the entire area around the pumping well. Pseudo-radial flow is evident on a log-log plot of drawdown versus time, or in derivative plots as flattened parts. At late time, when radial flow dominates, a classical method such as Theis or Jacob-Cooper can be applied. The drawdown response in observation wells located on the length of the fracture will be similar to the one in the pumping well, while the drawdown in observation wells located further away from the fracture will be delayed as a function of distance between the well and fracture.

Gringarten and Witherspoon (1972) determined the following solution for a pumping well intersected by a single, vertical fracture:

\[ s = \frac{Q}{4\pi T} F(u_{vf}, r') \]  \hspace{1cm} (2.10)

Where:

\[ u_{vf} = \frac{T t}{S x_f^2} \]  \hspace{1cm} (2.11)

\[ r' = \frac{\sqrt{x^2 + y^2}}{x_f} \]  \hspace{1cm} (2.12)

and

\[ S = \text{storativity of the aquifer, dimensionless.} \]
\[ T = \text{transmissivity (m}^2/\text{s).} \]
\[ x_f = \text{half length of vertical fracture (m).} \]
\[ x, y = \text{radial coordinate distances between observation well and pumped well, measured along x and y axes, respectively (m).} \]

Drawdown obtained in the observation well depends on the hydraulic characteristics of the well, but also on the geometrical relationship between the well and fracture. The aquifer geometry and the relationship between the pumping well and fracture are displayed as a 0.5 slope on the drawdown–time curve. A similar slope is observed for an
observation well if this well intersects the same fracture. At a distance beyond \( r' = 5 \text{ m} \) from the pumped well, the effect of the fracture will no longer be registered by changes in drawdown.

For a well-known geometry of the observation wells and the fracture, and for \( r' < 5 \text{ m} \), the Theis method can be applied to determine \( T \) and \( S \). In the present case, for an unknown geometry, a trial-and-error procedure must be used to select the most appropriate curve for observation wells from the three Gringarten-Whiterspoon set of curves.

Because the structural studies on the Gulf Islands suggest, there are numerous vertical and sub-vertical fractures, linear flow models, which account for linear flow, are perhaps most applicable in the current study region. Nevertheless, most of the data from the Gulf Islands region do not exhibit the entire behaviour expected for wells intersecting a fracture. This could be because wells are not necessarily located directly on a fracture, or that the wells are intersected by a network of fractures with each influencing drawdown independently. Therefore, even though single fracture models were used to analyze the data, the geometry of the aquifer and fracture could not be inferred because the drawdown curves did not perfectly match the standard curves. Thus, the single vertical fracture flow model was used only to estimate \( T \) and \( S \).

**Double–Porosity Concept**

Barrenblatt et al. (1960) described fractured formations as consisting of two overlapping media, each with its own properties. Flow through fractured / porous rocks is described as a re-equalization of the pressure differential in the fractures and unfractured rocks by the flow of fluid from porous media into the fracture. This kind of inter-porous flow is assumed to be in a pseudo-steady-state. Based on this theory, Bourdet and Gringarten (1980), Kazemi et al. (1969), and Warren and Root (1963), determined analytical solutions for \( T \) values for fractured rocks. A few sets of data were analyzed using double porosity models (Warren and Root, 1963; Kazemi et al., 1969), which consider an exchange of water between fractures and porous medium. These models proved not to be valid because, not a single double porosity drawdown curve
was observed in the entire database. This could be anticipated given the low primary porosity of the formations.

**Results**

Each aquifer test was integrated with geological and structural features, to assist in interpreting the effect of heterogeneities on the results. Site geological maps prepared by consultants, geological maps and air photos (Appendix B) were used to determine major structures and fractures that intersect or pass close by wells. This work was undertaken by Trevor Bishop (B.Sc., Department of Earth Sciences, SFU) as part of a summer research project. The aquifer tests graphs along with relevant air photos are provided in Appendix C (long duration tests) and Appendix D (short duration tests).

Aerial photos at a scale of 1:70 000 (see appendix B) were used to locate wells and determine relevant geologic information for each test. The location of each site is shown on a smaller scale scanned air photo of each island. Geological contacts and structural features are identified and used to integrate the test results with the geology of the area. Most of the tests are located near faults or fractures. This situation occurs because the consultants tend to locate the wells nearby a fault or fracture zone, in order to supply a client with a more reliable water supply. A consequence of using test data for wells completed near a fracture zone or fault is the bias that may be introduced (i.e., possibly higher parameter values).

The aquifers on the Gulf Islands exhibit mostly linear flow at early times - a steep slope on a log-log plot. Linear flow is followed by radial flow at mid to late times – flattened curve. This type of response indicates that initially the flow is associated with a single fracture or fracture zone intersected by the well, and as pumping continues, drawdown becomes more radial in nature, probably due to multiple fractures that act as an equivalent porous medium. Often radial flow is masked either by an extended linear flow period or by sudden changes in discharge rates.

In order to undertake numerical modeling, a conceptual model that characterizes the aquifer is required. In addition to identifying the hydraulic boundary conditions (e.g., recharge, rivers, ocean, etc.), the hydrostratigraphic units (or hydrostructural domains on the Gulf Islands) must be defined. Each unit is assigned a unique hydraulic conductivity (or permeability) and specific storage value.
Hydraulic conductivity (K) is a constant of proportionality defined as the volume of water that moves through a unit area of porous media, under a unit hydraulic gradient per unit time. One way to determine the hydraulic conductivity of fractured rocks is to conduct a packer test, in which case the K of each fracture is obtained directly. Alternatively, K can be derived from aquifer tests and calculated as shown in Equation 2.13.

\[ K = \frac{T}{b} \]  

(2.13)

where \( b \) is the saturated thickness of the aquifer.

Specific storage (\( S_s \)) is defined as the release of water from storage under unit decline of head from the compaction of the aquifer due to effective stress increase and expansion of water. \( S_s \) is calculated as:

\[ S_s = \rho g (\alpha + n\beta) \]  

(2.14)

Where:

\( \rho \) - density of water (kg/m\(^3\)),
\( g \) - acceleration gravity (m/s\(^2\))
\( n \) – porosity
\( \alpha \) – compressibility of aquifer (Pa\(^{-1}\))
\( \beta \) – compressibility of water (Pa\(^{-1}\))

In this case \( S_s \) was calculated from the storativity and saturated thickness of the aquifer.

\[ S_s = \frac{S}{b} \]  

(2.15)

The hydraulic tests described above are not packer tests, and therefore, they provide results in terms of transmissivity (T) and storativity (S). Thus, it is necessary to convert T and S into hydraulic conductivity (K) and specific storage (\( S_s \)), respectively. In porous media the saturated thickness (b) is represented by the length of the well screen or the open hole length. In fractured rocks, it is hard to approximate the saturated thickness of the aquifer because it depends on the density and width of fractures. A
method to approximate $b$ is to sum the apertures of the fractures along the entire length of the well. This method was not possible in this study because of the uncertainty in fracture location and aperture along each well tested. Rather, it is assumed that $T$ values characterize radial flow over the entire open interval (between base of well case and bottom depth of the well). However, this approximation of the saturated thickness is expected to underestimate hydraulic conductivity values by perhaps an order of magnitude or more.

A literature review provided the following range of $K$ values for sandstone and mudstone rocks (Table 2.2).

**Relation to Geology**

The geology of the Gulf Islands is represented by the fractured sedimentary rocks of the Upper Cretaceous Nanaimo Group (Figure 2.12). The Nanaimo Group consists of an alternating and intertonguing sequence of sandstones, mudstones, siltstones and conglomerates deposited in a marine environment. The most recent review of the Nanaimo Group nomenclature was done by Mustard (1994), who identified eleven lithostratigraphic formations, and interpreted these as submarine fans deposits. Formations alternate between sandstone-dominant and mudstone-dominant, with transitional zones where bed thickness is reduced. As a consequence of the transition zones, a strict separation between formations is hard.

A regional deformation event during the Late Eocene resulted in northwest crustal contraction that is defined by a southwest vergent fold and thrust belt – Cowichan Fold and Thrust system (England and Calon, 1991). The region was later uplifted and eroded during the Late Neogene – uplift of the Coast Belt (Mustard, 1994). It is suggested that these deformational events likely developed secondary porosity (i.e., fractures), thus increasing hydraulic conductivity by perhaps a few orders of magnitude. The sandstone-dominated formations are observed to have a low fracture density, while the mudstone-dominant formations are more highly fractured. Interbedded zones have the highest density of fracturing, because they tend to be more thinly bedded.
Figure 2.12 Geology of Saturna Island represented by Nanaimo Group (geology after Mustard, 1994).
Representative K and S\textsubscript{s} values for each formation, where test data are available, are provided in the following sections, along with a brief description of the stratigraphy, as summarized from Mackie (2002). The K and S\textsubscript{s} values were derived from tests conducted throughout the entire Gulf Islands region, and are assumed to be representative of each formation. Mackie (2002) also determined fracture intensity as the number of fractures per unit length for almost all Nanaimo Group formations. These data are also summarized for each formation where available.

**Protection Formation**

The Protection Formation is a 60 m sequence of medium to thick-bedded sandstone, interlayered with conglomerate and siltstone. Generally, the structure is massive or cross-rippled at the top. Interbeds consist of normal graded pebbly conglomerate scouring into sandstone, which grades to thin beds of silty mudstone. On Saturna Island, the Protection Formation is exposed as a 150 m wide band on southern part of the island from Crocker Point to Taylor Point (Figure 2.12).

Hydraulic conductivities and specific storage (Table 2.3) values are relatively low, but there is also a low level of confidence in these results because only one long duration test, including a pumping well and two observation wells, was analyzed for this formation.

**Cedar District Formation**

The Cedar District Formation is present as a sequence of massive mudstone containing abundant calcium carbonate concretions and massive mudstone with interbeds of siltstone and sandstone. It is a moderate to deep marine environment facies, which coarsens and thickens upwards to medium sandstone-mudstone couplets. The formation is present on Saturna Island between Narvaez Bay and Lyall Harbour (100 m in width), and on southern part of the island (150 m in width) (Figure 2.12). Fracture intensity for this formation is 1.22; low by comparison to the other formations outcropping on Saturna Island.

K and S\textsubscript{s} values were obtained from two tests conducted in Cedar District Formation (Table 2.4)
de Courcy Formation

The de Courcy Formation consists of thick-bedded layers of medium to coarse-grained sandstone with thick lenses of conglomerate, interbedded with sandstone, siltstone and mudstone. The thick sandstones are normally separated by thin-bedded fine arenite and massive siltstone or mudstone. The de Courcy was deposited in middle to upper submarine fan. The formation outcrops on Saturna Island between Winter Cove and Narvaez Bay, and on south-central part of the island (Figure 2.12) where it is approximately 300 m in width.

The hydraulic parameters for this formation are slightly higher than those of other sandstone formations (Table 2.5). Fracture intensity is higher at 3.14. K and S values could not be determined directly from T and S results, as there are no data about the depth of well or casing length. However, an average well depth of 100m was assumed, based on all the depths in this study, and used to calculate these values.

Northumberland Formation

The Northumberland Formation consists of thick beds of grey mudstone and siltstone, interbedded with sandstones. Mudstone layers occur mainly as massive beds, but with discontinuous, thin isolated layers of sandstone. The formation is approximately 150-180 m wide in the Brown Ridge area and Winter Cove, and it extends to the southeast side on Warburton Pike and Mount Fisher (Figure 2.12).

No tests were conducted in this formation, and no fracture measurements were made. Therefore, hydraulic parameters are estimated based on the parameters of similar mudstone-dominant formations and from literature values.

Geoffrey Formation

The Geoffrey Formation consists of thick to medium bedded coarse arenite to boulder conglomerate, with minor finer interbeds of sandstone. It outcrops in Brown Ridge area and on Winter Cove and along the entire north edge of the Island through to East Point (Figure 2.12). The formation was deposited in middle to upper part of submarine fan environment. It is massive, and attains a width of 300 m on the north side
of the island. A fracture intensity of 2.65 is reflected in the higher values of T compared with formations with a lower fracture intensity.

Values for hydraulic parameters (table 2.6) are susceptible to error, because most of the tests were conducted simultaneously with some other nearby tests. Because no information on the depth of the wells or on the length of the casing was available, a depth of 100 m was considered, by averaging the thickness of Geoffrey Formation on Saturna Island.

**Spray Formation**

The Spray Formation consists of a sequence of thin beds of grey mudstone and siltstone, with isolated thin beds or interbeds of sandstone deposited as a part of lower and middle fan in distal areas. It is a graded sequence, with rare intercalations of matrix supported conglomerate. On Saturna Island, the Spray formation occurs only at the northern edge of the island, as a thin outcrop near Tumbo Channel, where it is about 50 m wide. Fracture intensity varies between 11.3 and 0.33.

Transmissivity varies in between $10^{-4} - 10^{-7}$ m$^2$/s (Table 2.7).

**Summary of Hydraulic Parameters**

Hydraulic parameters measured in fractured rocks are typically very different from one location to another, and depend on the distribution and types of fractures, which may act as either paths for groundwater flow or barriers to flow (in the case of faults). No trends could be identified in the results of T or S to relate the resulted values to the duration of the test or to the lithology. However, a general pattern for all tests indicates linear flow for early times, radial flow at mid times and hydraulic barriers at late time. The results of long duration tests are considered to be more reliable, because the aquifer is tested for a longer period of time and the flow can attain the radial behaviour, as well as capture the effects of boundaries.

As shown in Figure 2.13, T results obtained using long duration tests (bubbles on the first row) are slightly lower than those obtained using short duration tests (bubbles on the bottom row), possibly due to the fact that when the aquifer was stressed for a short
period of time, it did not attain a radial flow. Therefore, the influence of the fracture dominated the well response for the short duration tests, yielding higher values for T. T values shown in Figure 2.13 were derived from Theis analyses.

The Spray and Cedar District formations are mudstone dominant, while the Geoffrey, de Courcy and Protection formations are sandstone-dominant. Based on these lithology types, there are no observable trends that would suggest that one or the other type of lithology has a higher permeability as was anticipated from the fracture mapping conducted in the region. In fact, the Geoffrey and de Courcy formations have the highest T values (almost an order of magnitude higher than the other formations). It is suggested that because these tests were conducted in wells situated in close proximity to major fracture zones that the results may be biased (Allen et al. 2003). In fact, most of the wells tested are located nearby mapped faults or fracture zones. Therefore, the T and S values, and consequently, the K values may overestimate the relatively unfractured aquifer formation away from the influence of the fault. This presents a potential problem for selecting the appropriate K value for the aquifer. For modeling purposes, the full range of values obtained will to be tested to represent the aquifer formations.

The T values are comparable for all methods used for analysis, and in particular, the Theis and Recovery methods. These favourable results increase the confidence in the values obtained, because recovery method is independent of pumping rate variations that could affect the pumping test data. The confidence in K values increases with the number of tests conducted in each location, as shown in Figure 2.14. Therefore, the most reliable values are those for the Geoffrey Formation.
Figure 2.13  Comparison between T results obtained from long and short duration tests data, using Theis and Recovery methods

Increase in grade of confidence on K value for each formation, based on the number of tests used to determine an average k

Note: The size of bubbles is determined by the average hydraulic conductivity of each formation

Figure 2.14 Degree of confidence in K values, based on the number of tests conducted in each formation.
<table>
<thead>
<tr>
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<tbody>
<tr>
<td>Sandstone</td>
<td>$10^{-5} - 10^{-3}$</td>
<td>$10^{-3}$ - $1$</td>
<td>$10^{-5} - 10^{-3}$</td>
</tr>
<tr>
<td>Mudstone</td>
<td>$10^{-8} - 10^{-5}$</td>
<td>$10^{-7}$</td>
<td>$10^{-7} - 10^{-5}$</td>
</tr>
<tr>
<td>Fractured rock</td>
<td>$10^{-4} - 10^{-2}$</td>
<td>$0 - 3 \times 10^{2}$</td>
<td>$10^{-4} - 10^{-2}$</td>
</tr>
</tbody>
</table>

Table 2.2 Aquifer test distribution by geologic formation.

<table>
<thead>
<tr>
<th>Method</th>
<th>Theis</th>
<th>Jacob 1</th>
<th>Jacob 2</th>
<th>Gringarten</th>
<th>Recovery</th>
</tr>
</thead>
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<tr>
<td></td>
<td>K (m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
</tr>
<tr>
<td>Averages</td>
<td>2.25E-07</td>
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<td>1.45E-07</td>
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Table 2.3 Hydraulic parameter values - Protection Formation.
<table>
<thead>
<tr>
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<th>Theis</th>
<th>Jacob 1</th>
<th>Jacob 2</th>
<th>Gringarten</th>
<th>Recovery</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hydraulic Parameter</td>
<td>K (m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
</tr>
<tr>
<td>Averages</td>
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Table 2.4 Hydraulic parameter values - Cedar District Formation.

<table>
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<th>Jacob 2</th>
<th>Gringarten</th>
<th>Recovery</th>
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</thead>
<tbody>
<tr>
<td>Hydraulic Parameter</td>
<td>K (m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
</tr>
<tr>
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Table 2.5 Hydraulic parameter values - De Courcy Formation.
<table>
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<th>Jacob 2</th>
<th>Gringarten</th>
<th>Recovery</th>
</tr>
</thead>
<tbody>
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<td><strong>Hydraulic Parameter</strong></td>
<td>K (m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
</tr>
<tr>
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<td>4.71E-08</td>
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<td>1.96E-06</td>
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Table 2.6 Hydraulic parameter values - Geoffrey Formation.

<table>
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<tr>
<th>Method</th>
<th>Theis</th>
<th>Jacob 1</th>
<th>Jacob 2</th>
<th>Gringarten</th>
<th>Recovery</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Hydraulic Parameter</strong></td>
<td>K (m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
<td>Ss</td>
<td>K(m/s)</td>
</tr>
<tr>
<td>Averages</td>
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<td>N/A</td>
<td>5.81E-07</td>
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</table>

Table 2.7 Hydraulic parameter values - Spray Formation.
CHAPTER 3
SALTWATER INTRUSION AND DENSITY–DEPENDENT FLOW:
USGS, SUTRA CODE

Saltwater intrusion refers to the movement of salty water into aquifers, and the resulting contamination of freshwater resources. The magnitude of saltwater intrusion is typically expressed as a function of chloride concentration. Concentrations higher than 150 mg/l Cl result in poor drinking water quality, and therefore, are said to reflect the occurrence of saltwater intrusion.

Under natural conditions in a coastal aquifer, groundwater discharges towards the sea or ocean and the freshwater flows above the stagnant mass of saltwater. The equilibrium can be affected by natural and anthropogenic factors, as discussed in Chapter 1. Transport occurs primarily by advection, but ions also move due to molecular diffusion. Therefore, groundwater flow is controlled by the hydraulic gradient, while the transport of solutes is determined by the groundwater velocity and the gradient of concentrations. Depending on the magnitude of the concentration gradient, the freshwater-saltwater interface will either be sharp (narrow interface) or transitional (more disperse) in nature.

Density Dependent Flow

The ease with which groundwater flows is controlled by two factors: 1) the geological medium and 2) the properties of the fluid, including the fluid density and viscosity. While heterogeneity, at a variety of scales, is a consequence of spatial variations in the geologic media, fluid density and viscosity are affected by both temperature and the amount of dissolved solids. In freshwater aquifers, the temperature and the concentration of total dissolved solids do not vary significantly, therefore, the fluid properties remain constant, and the variations in flow are the result of variations in the geologic medium and the hydraulic gradient. In high- or low-temperature environments or in aquifers that have high concentrations of total dissolved solids, the fluid properties may change spatially. This can result in density-dependent flow.
The mechanism for density-dependent flow is determined by the density of fluids. When the density of one fluid is greater than the ambient water density (for example, by the addition of significant amounts of dissolved solids) free convection can determine the transport of solute over large distances. In the case of a coastal aquifer, fluid density can vary over short distances, particularly in the vicinity of the saltwater - freshwater wedge. This results in flow instabilities, and ultimately, modifications to the position and width of the freshwater-saltwater interface.

Density is a function of pressure, temperature and concentration of dissolved solutes. An equation of state expresses the linear relationship between chloride concentration and seawater density:

\[ \rho(C) = \rho_f (1 + \alpha C/C_s) \]  

Where:
\[ \rho(C) \] = density of groundwater (kg/m³);
\[ \rho_f \] = density of freshwater (kg/m³);
\[ \alpha \] = relative density difference;
\[ C \] = chloride concentration or chlorinity (mg Cl/L); and
\[ C_s \] = reference chloride concentration (mg Cl/L).

Saltwater intrusion is defined by a concentration of total dissolved solids (TDS) greater than 400-500 mg/L. The interface is usually approximated by the 50% saltwater isochlor. The saltwater-freshwater interface depends on both variability of saltwater density and on mixing of salt and freshwater.

There are two major conceptual models for representing the saltwater-freshwater interface. The two approaches are related to the shape of the interface: sharp and transitional. When the transition zone between fresh and saltwater is much thinner than the rest of the freshwater body, then it can be approximated as a line between the two types of water (i.e., the sharp interface case). This case occurs rarely and is associated with sand-dune areas or coral islands, where the freshwater lens evolves by natural groundwater recharge (Oude-Essink, 2001). As most systems are not in hydrostatic equilibrium, the transitional interface approach is more suitable. The transition zone in a coastal aquifer varies from one place to another, and it is
represented by gradual zones of fresh to brine to saline water resulting from dispersion, diffusion and physical and chemical processes. The width of the diffuse zone and the amplitude of saltwater intrusion depend on various factors, such as: type of aquifer, aquifer geology and geometry, tidal effects, rate of recharge, rate of evaporation, seawater density, rate of withdrawal, the hydraulic characteristics of geological units and their geometry, hydrogeological boundaries, barometric pressure, earth tides, chemical changes and wave action (Walton, 1970).

On a small island aquifer system, such as is being considered in this research, a less dense freshwater body is underlain by a heavier mass of saltwater (Figure 3.1). Henry (1959) stated that the movement of freshwater towards the sea acts as a barrier to seawater intrusion into aquifer. The freshwater floats on top of saltwater and the interface has the shape of a lens. Under normal conditions, both waters are in equilibrium. The freshwater lens equilibrium can be disturbed by water extraction, which can result in upconing and/or the movement of the interface inland.

**Location of Saltwater-Freshwater Interface – Ghyben-Herzberg**

First solution for the position of the saltwater – freshwater interface was proposed independently by Badon-Ghyben (1888) and Herzberg (1901), and is commonly known as the Ghyben-Herzberg solution. This approximation assumes that saltwater and freshwater are immiscible fluids. The analytical solution provides a position of the sharp saltwater – freshwater interface in a coastal aquifer under hydrostatic conditions in homogeneous coastal aquifer, where the flow is perpendicular to the coast.

It is assumed that:

Pressure saltwater = Pressure freshwater

This condition of hydrostatic balance is expressed as:

\[ \rho_s \times H \times g = \rho_f \times (h + H) \times g \iff \rho_s \times H = \rho_f \times H + \rho_f \times h \]  

(3.2)

\[ h = \frac{\rho_s - \rho_f}{\rho_f} \iff h = \alpha H \]  

(3.3)

Where \( H \) = depth below sea level to a point on the interface;
Figure 3.1 Saltwater-freshwater interface on a oceanic island.
\[ H = \text{height of freshwater column}; \]
\[ \rho_s = \text{density of saltwater (typically 1.025 g/cm}^3); \]
\[ \rho_f = \text{density of freshwater (typically 1.000 g/cm}^3). \]

If typical values of \( \rho_s \) and \( \rho_f \) are used, the Ghyben–Herzberg solution gives the solution that the depth of the interface is 40 times the difference of the height of water table above mean sea level (or \( h = 40H \)).

This solution is viable only under following conditions:

- Homogeneous aquifer,
- Saltwater is at rest,
- Flow in the aquifer is only horizontal (vertical flow is negligible),
- Hydrodynamic dispersion is negligible.

The Ghyben-Herzberg solution gives only an approximation of the position of the interface, which is erroneously assumed to be positioned in the outflow zone as illustrated in Figure 3.2. This figure shows the Ghyben-Herzberg assumptions and the difference between the real and calculated positions of the interface.

Later, Glover (1964) published an analytical solution for a sharp saltwater-freshwater interface for steady-state flow in a homogeneous, confined aquifer. This solution is a modification of the Ghyben–Herzberg solution, and considers the freshwater flow discharging to the sea. In this solution, the position of the interface is a function of hydraulic conductivity and of the flow per unit width of coast line. The interface has a parabolic shape, and is shifted in the \( x \)-direction and is located at (Figure 3.3):

\[
H = \sqrt{\left( x - \frac{q_0}{2k} \right) \left( -\frac{2q}{k} \right)} \quad (3.4)
\]

Where:

- \( H \) = depth of interface;
- \( x \) = distance inland from the shore (\( x = 0 \) at the shoreline);
- \( K \) = hydraulic conductivity (m/s);
Figure 3.2 The Badon Ghyben-Herzberg principle, a fresh-saltwater interface in an unconfined coastal aquifer.

Figure 3.3 Saltwater-freshwater interface in a confined aquifer, Glover solution (1964) freshwater discharging towards the ocean.
\[ q_0 = \text{flow per unit width through the aquifer at the coastline} \]

\[ \alpha = \text{relative density difference} \frac{\rho_f - \rho_s}{\rho_f} . \]

Numerical studies were conducted by Henry (1959), Shamir and Dagan (1971), Bear and Dagan (1964), and Bear (1972), which provide solutions based on the Dupuit–Forchheimer approximation. These solutions are based on the fact that flow in regional groundwater models is essentially horizontal and that at the interface the weight of column of freshwater is balanced by the weight of column of seawater extending from sea level to the interface position, disregarding the effect of diffusion and dispersion. Fetter (1972) approximated the position of the interface in a two-dimensional steady-state unconfined aquifer, and later Anderson (1976) calculated the position of the intrusion for a one-dimension strip-oceanic island, and developed a formula that predicts the depth of the interface. Cox (1951) described the effect of tides on atoll islands, which represent a dual aquifer system, subject to vertical and lateral tides.

**Location of Saltwater-Freshwater Interface - Transition Zone**

In reality, saltwater and freshwater are miscible fluids and the interface has a more complex shape. The two types of waters merge into a transitional zone, with varying fluid concentrations and densities increasing from the freshwater to the saltwater body. Fluid density varies gradually from higher values at the sea boundary to lower values inland.

Brown (1925) noted that the interface is more a transition zone rather than a sharp interface, and fluctuates due to eustatic sea level, precipitation, or pumping. The fluctuations cause dispersion: "the process by which two miscible liquids interfuse around their boundary when hydraulic flow causes the boundary to move". Later, Cooper (1959) observed that under certain conditions, the interface between freshwater and saltwater is wide, and the sharp approach is not valid. He introduced the transitional zone approach, which considers a dynamic equilibrium and miscibility between fluids that produces a transition zone.
In coastal aquifers that have been subjected to numerous transgressions and regressions, hydrodynamic dispersion is an important factor for determining the magnitude of saltwater intrusion. Hydrodynamic dispersion \( D_h \) is defined by two components:

\[
D_h = D + D_m
\]

Where mechanical dispersion \( D \) is determined by fluid velocity variations, results in mixing between groundwater and contaminant, and consequently, dilution of the contaminants. Molecular diffusion \( D_m \) is due to difference in concentration that results in a movement of ions or molecular species dissolved in water (in saltwater molecular diffusion is approximately \( 10^{-9} \text{ m}^2/\text{s} \)). In natural groundwater systems, hydrodynamic dispersion depends mostly on mechanical dispersion; molecular diffusion has only a minor effect.

The interface width is a function of the aquifer hydraulic parameters, the density of saltwater, the transport parameters (e.g., dispersivity), aquifer geometry and anisotropy, exploitation and distribution of pumping wells, and recharge rate. Tidal activity also increases the dispersion effect. In these cases, a variable density model with solute transport must be used to determine the movement and the position of interface.

Henry (1964) investigated the role of diffusion and dispersion in creating a wide transition zone in variable density flow, by simulating the boundary conditions at the sea side of a coastal aquifer. In this case, the saltwater - freshwater interface is represented by a transition zone as shown schematically in Figure 1.1. By considering a homogeneous, isotropic, confined and rectangular coastal aquifer, Henry derived analytical solutions for stream functions and concentrations as Fourier series. This example has become standard solution used to verify numerical codes that simulate density-dependent flow.

Of specific relevance to this study is the work by Hamza (2000), who concluded that the distribution of chloride concentrations is sensitive to longitudinal dispersivity, and also depends on travel distance. The shape and the degree of intrusion depend on the geological features and on the hydraulic conductivities. In a layered aquifer, saltwater intrudes more or less as the hydraulic conductivities increase or decrease, respectively. Under such conditions more complex solutions, which consider both density-dependent
flow and solute transport, should be applied to determine the magnitude of saltwater intrusion.

Therefore, saltwater intrusion is a problem of density-dependent groundwater flow and solute transport as solutes affect the density of groundwater (Voss, 1984). Considering that the fluids are miscible and the flow is driven by the solute concentration coupled to the fluid density, it is necessary to determine simultaneous solutions of the variable density fluid flow equation and the advection-dispersion equation to describe the salt transport. This is expressed mathematically as:

\[
\frac{\partial C}{\partial t} = \frac{\partial}{\partial x_i} \left( \frac{D_{ij}}{\partial x_j} \right) - \frac{\partial}{\partial x_i} (CV_i) + \frac{(C - C')W}{n_e} + \frac{\Psi}{n_e} \tag{3.5}
\]

Where:

- \( C = \) concentration of dissolved solutes (kg/m³);
- \( t = \) time (s);
- \( D_{ij} = \) hydrodynamic dispersion tensor (m³/s);
- \( V_i = q_i / n_e \) (effective groundwater velocity in \( x_i \) direction) (m/s);
- \( C' = \) concentration of dissolved solids in a source (kg/m³);
- \( W = \) general term for a source (s⁻¹);
- \( n_e = \) effective porosity of the medium (unitless);
- \( \Psi = \) chemical reaction of source per unit volume (includes equilibrium-controlled sorption or exchange and first-order irreversible rate reactions) (kg/m³s).

At steady-state it is assumed that mass is conserved in a system and it is described by the following partial differential equation:

\[
\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \text{ (rate of change of hydraulic head is constant with time)} \tag{3.6}
\]

The constitutive equations for coupled saltwater–freshwater flow in an anisotropic non-homogeneous media are defined by Bear (1972):
\[ q_x = -\frac{k_x}{\mu} \frac{\partial P}{\partial x} \]  (3.7) where:

\( q \) = Darcian flux (m/s);
\( k \) = intrinsic permeability (m²);
\( \mu \) = dynamic viscosity of water at a point (kg/ms);
\( P \) = pressure (kg/ms²);
\( \rho \) = fluid density (kg/m³);
\( g \) = gravitational acceleration (m/s²).

Saltwater intrusion problems are solved using several solution methods:

- Analytical solutions – solve equations by specifying heads and discharge rates to determine exact solutions of interface.
- Finite difference method – based on Taylor series, and is easier to code and usually faster to solve.
- Finite element method - good performance for problems that require a wide range in element size, and also give a good representation of boundary conditions.
- Analog models - investigate the behaviour of density-dependent groundwater flow, and are applied to study groundwater flow in specific situations. These are translated into equations and flow constants (Hele-Shaw model or electrical model).
- Random walk method - uses particle tracking method.

**Sutra Code**

Numerical models are now standard methods for calculating present and future configurations of the saltwater interface. Codes to determine different configurations of saltwater intrusion as a sharp interface or as a transitional zone are now available:
- SUTRA (Voss, 1984) – 2D finite element model that simulates saturated-unsaturated, fluid density-dependent flow;
- HST3D (Kipp, 1986) - 3D finite element code the simulates salt and heat transport;
- SWICHA (Huyakorn et al., 1987) - for coupled flow and solute transport in porous media;
- METROPOL (Sauter et al., 1993) - that simulates density driven flow and contaminant transport;
- MOC3D (Konikow et al., 1996) - suitable for transient three-dimensional groundwater flow in large-scale hydrogeologic systems where non-uniform density distributions occur.

For the purposes of this thesis, the SUTRA code (Voss, 1984) is used to simulate the effect of heterogeneities present as layers on Saturna Island, and to model the saltwater-freshwater interface movement over the last 12,000 years. SUTRA (Saturated-Unsaturated Transport Model) is a finite element code that simulates saturated-unsaturated, fluid density-dependent flow. The 2D version can be used for both solute and energy transport, in either aerial or cross-section view. SUTRA is also available in 3D (released in 2003). SUTRA is written in Fortran 77, employs a finite element discretization based on the Galerkin method, and produces stable solutions with a relatively coarse mesh. SUTRA has the advantage of an irregular mesh that properly accommodates the island outline, it has a proper temporal discretization, and it is stable.

This code has been used in several case studies to simulate the transition zone. These include studies involving flow and transport in a coastal aquifer in Oahu, Hawaii (Souza and Voss, 1987); evaluation of groundwater flow in fractured rocks (Andersson et al., 1991), use for evaluating a remediation scheme (Wagner and Gorelick, 1987), and testing climate change impacts on saltwater intrusion in India (Bobba et al., 2000).

Finite-element methods use assumed functions of the dependent variables and parameters to evaluate equivalent integral formulations of the partial differential equations. The rate of flow of water through a porous media is related to the properties of the water, the properties of the porous media, and the gradient of the hydraulic head; a solution for the transient flow can be derived from Darcy's law. The simulation of
seawater intrusion requires the solution of partial differential equations that describe “conservation of mass of fluid” and “conservation of mass of solute” (Voss, 1984).

Groundwater flow is calculated by solving the fluid mass balance equation:

\[
\frac{\partial (\varepsilon \rho \phi)}{\partial t} = - \nabla \cdot (\varepsilon \rho V) + Q_p
\]  

(3.8)

where:

\( \varepsilon(x,y,t) \) = a dimensionless porosity;
\( \rho(x,y,t) \) = fluid density (kg/m³);
\( V(x,y,t) \) = average fluid velocity (m/s);
\( Q_p(x,y,t) \) = fluid source or sink (kg/m³ s), with \( x, y \) – coordinate variables (m), \( t \) – time (s).

This equation can be expressed in terms of pressures and concentrations as:

\[
(\rho S_{op}) \frac{\partial p}{\partial t} + \left( \varepsilon \frac{\partial \rho}{\partial C} \right) \frac{\partial C}{\partial t} - \nabla \left[ \frac{\rho k}{\mu} \right] \cdot (\nabla p - p g) = Q_p
\]  

(3.9)

where:

\( S_{op} \) = specific pressure storativity that is a function of porous matrix compressibility \( \alpha \), (m·s²/kg) and fluid compressibility \( \beta \), (m·s²/kg);
\( C(x,y,t) \) = solute concentration expressed as mass fraction;
\( k(x,y,t) \) = permeability;
\( \mu(x,y,t) \) = fluid viscosity;
\( p \) = fluid pressure;
\( g \) = gravitational acceleration.

Transport is solved by the solute mass balance equation:

\[
\varepsilon \rho \left( \frac{\partial C}{\partial t} \right) + \varepsilon \rho V \cdot \nabla C - \nabla \left[ \varepsilon \rho (D_m I + D) \cdot \nabla C \right] = Q_p \left( C^* - C \right)
\]  

(3.10)

where:

\( D_m \) – is molecular diffusivity of a solute in solution in a porous medium (m²/s);
I - identity tensor;

\( C^*(x,y,t) \) - concentration of solute as a mass fraction in the source fluid (kg salt/kg freshwater);

\( D(x,y,t) \) – dispersion tensor (m²/s), which in 2D is given by ,

The fluid density \( \rho(x,y,t) \) is defined as a linear function of concentration following:

\[
\rho = \rho_0 + \frac{\partial \rho}{\partial C}(C - C_0) \quad (3.11)
\]

where:

\( \rho_0 \) is fluid density when \( C = C_0 \);

\( C_0 \) is base solute concentration; and

\( \frac{\partial \rho}{\partial C} \) is a constant coefficient of density variability.

The principle of conservation of mass requires that the net mass of solute entering or leaving a specified volume of aquifer during a given time interval must equal the accumulation or loss of mass stored in that volume during the interval.

Modeling for this thesis was performed using the US Geological Survey's graphical-user interface for SUTRA (SUTRA GUI) as a plug-in extension (PIE) to Argus ONE (Argus Interware software). The results were visualized using SUTRA Plot 2D3D-2(β version), a USGS post-processing graphic utility (Souza, 2003). SUTRA simulations require input data in terms of pressure and salinity distribution with depth, hydraulic parameters, dispersivity and recharge estimates, as well as topographical profiles. The output files allow an interpretation of saltwater intrusion magnitude in terms of salinity and pressure distribution.

Conceptual Model

A numerical model is a mathematical representation of a physical system (in this case an aquifer), that, using governing laws of groundwater movement can reproduce
aquifer behaviour over time. A conceptual model is a tool that ensures the relationship between components of a system while simplifying field conditions. It represents the input in numerical modeling. The conceptual model aims to reproduce the field parameters in terms of geology, water quality, and recharge in relationship with rivers, water levels, hydraulic parameters, or pumping. Simple and well-defined models, which include only key data as boundary conditions and aquifer properties, are the best way to start numerical modeling. Complexity is added afterwards.

The main objectives of the present conceptual model are to:

1. Determine the role of heterogeneities on the configuration of the saltwater–freshwater interface.
2. Test the theory of saltwater-freshwater interface movement during the last 11,000 years. Specifically, the rate of return of the freshwater-saltwater interface to equilibrium following submergence of the island will be simulated.

For the second objective, the problem to be tackled is:

Assuming representative equivalent K values for the aquifer (as derived from simulations conducted as part of objective 1) and a rate of recharge, how long does it take for the freshwater-saltwater interface to attain an equilibrium position assuming that the island was initially filled with seawater for a period of about 1,000 years roughly 10,000 years ago? By knowing the current distribution of chloride in a test area (East Point), a model can be calibrated to constrain the estimates of K and recharge. Using these estimates at an island-wide scale will then allow simulations to be conducted to constrain the time required to move the interface and answer the above question. Hydraulic parameters that play a key role in the structure of a model were described in the previous chapter; other factors are described in the following sections.

Model Domain and Physiography

As a consequence of amount and distribution of data, three cross-sectional models, and consequently, three conceptual models are constructed for this area to test both the effect of heterogeneities on interface configuration and saltwater interface movement during the last 10,000 years. Cross-section locations are as shown in Figure...
3.4. Two of the cross-sections are located on East Point peninsula, the southernmost part of the island, where a large amount of data is available. The construction and calibration of these two cross-sections will provide a constraint of the parameters to be used for the longer cross-section across the entire island. The thickness of the model is 1m, as all the models are 2-D cross-sections.

The aquifer profile controls the groundwater flow, along with the boundary conditions. For each model, a cross-section was constructed using a topographic map of the island, as well as bathymetric maps (Haro Strait, Boundary Pass and Satellite Channel Map, 1996). In the cross-sectional models, topographical data and bathymetric data constrain the shape of the model domain. The highest elevation on the island is 410 meters; Mt. Warburton Pike. On the northeast side, the coastline is gently sloping, and on the southwest side it is mostly rugged, consisting of wave-cut cliffs and steep promontories. The base of the model was set to a depth sufficient to represent groundwater flow and saltwater intrusion patterns, to have a good representation of the saltwater–freshwater interface, and to include heterogeneities that reside in layers with different permeability (k) values as a consequence of fracturing. The depth of the base of the model is somewhat proportional to the length of the model, with longer cross-sections requiring a deeper base. As indicated in Figure 3.5 (across East Point peninsula) and Figure 3.6 (along strike of East Point peninsula) geological cross-sections have a depth of 300 to 350 m, respectively. The cross-section across the entire island, as shown in Figure 3.7, is deeper (700 m) due to topographic variations on the order of 0 and 410 m. In all models, the depth of the cross-section is chosen such that groundwater flow will be horizontal (parallel to the bottom of the domain outline).

**Geology**

In terms of geological features, Saturna Island consists of a thin veneer of glacial sediments, underlain by Late Cretaceous, Nanaimo Group sedimentary rocks that were deposited on Paleozoic to Jurassic basement rocks, of uncertain depth. The glacial sediments are not important as a groundwater supply, and therefore, are not considered further. However, it is recognized that these deposits may play a role in groundwater recharge. Six geologic formations are exposed on Saturna Island, including: Protection Fm., Cedar District Fm., de Courcy Fm., Northumberland Fm, Geoffrey Fm., and Spray Fm.
Figure 3.5 Geological cross-section across East Point peninsula (A-A'). Northumberland Formation is extrapolated to the base of the model domain.
Figure 3.6 Geological cross-section along strike of East Point peninsula (B-B') Northumberland Formation is extrapolated to the base of the model domain.
While there are no precise data on the thickness of each formation on Saturna Island, for the purposes of the current study, thicknesses were approximated from geological data from studies on similar islands (e.g., Hornby Island by Katnick and Mustard, 2001); from geological studies of the region (Mustard, 1994) and from geophysical logs (Abbey, 2000). Because of the uncertainty in basement depth, the lowest known formation has been extrapolated to the base of the model domain as illustrated in Figure 3.7.

Of direct relevance to this study, is the development of preferential pathways for the fluid flow and intrusion of saltwater. Because the geology is not represented by a layer-cake type of stratigraphy, rather by interlayers, the saltwater interface is speculated to have a complex appearance. Therefore, for the purposes of this study, the role that fractures play at a small scale will be down-played. Rather the larger-scale dynamics of the wedge will be simulated at a regional scale by assuming that the aquifer can be represented as an equivalent porous medium as discussed in the previous chapter.

**Boundary Conditions**

Boundary conditions are physical or chemical constraints that apply to the model domain, which represent or express the physical/chemical boundaries of the aquifer that is modelled. Three types of boundary conditions are commonly specified in numerical modeling: Dirichlet or specified heads (pressure), Neuman or specified flux, and Cauchy or head dependent flow. The Sutra code requires specification of pressure rather than hydraulic heads (Voss, 1999).

**Dirichlet type boundary conditions** require that pressure be specified for the sea boundaries. Pressure varies linearly with depth according to

\[ P = \rho_w g d \]  

(3.12)

Where \( \rho_w \) = density of water, \( g \) = gravitational acceleration, and \( d \) = depth below the water table. Thus, the boundary is a specified pressure as a function of water density.

Therefore, specified pressures are assigned where the aquifer is in contact with the ocean according to the following equation:
Figure 3.7 Geological cross-section across entire island (C-C'). Protection Formation is extrapolated to the base of the model domain.
\[ \rho(C) = \rho_f + \frac{\partial \rho}{\partial C}(C - C_0) \quad (3.13) \]

\( \rho_f \) = density of fresh water (kg/m³),

\( C \) = mass fraction of total dissolved solids. The mass fraction of seawater, \( C_0 \), equals 0.0357 kg (dissolved solids) / kg (seawater),

\[ \frac{\partial \rho}{\partial C} \] = a constant value of density change with concentration, e.g. 700 (kg/m³) or kg² (seawater) / kg (dissolved solids) m³.

Specified pressure boundaries are assigned as shown in Figures 3.8, 3.9, 3.10 on the ocean sides, down to the depth measured on the bathymetric maps.

In addition to specified pressure boundaries being used to represent the physical boundary conditions for the ocean, these model boundaries also require some representation of chemical concentration because ocean water has a different concentration than freshwater. Thus, the specified pressure nodes in the model are assigned a solute concentration. A specified concentration boundary condition is not used in the case of saltwater intrusion because the salt is delivered to the model as a consequence of advection and dispersion of a moving water mass as well as diffusion.

The density differences of the supplied water at this boundary influence the flow regime. This is in contrast to the example of a salt dome which supplies an endless source of constant concentration solute that simply becomes dissolved in the water. Conductivity measurements taken on different parts of the island suggested that solute concentration differs along the two sea boundaries (north and south edges). Concentration on the north shore is lower due to mixing with fresh waters of the Fraser River. On the north edge, water is considered to have TDS= 20,000mg/L, and on the south side TDS= 35,000mg/L (after Stuyfzand, 1993).

**Neumann boundary conditions** (equal to the rate of recharge) are applied at the water table. The water table was estimated from available data or estimated where no measurements were available. In reality, the water table is at some depth below ground surface, and this depth may be, in fact, significantly deeper, particularly in the central portion of the island. Previous studies have indicated that “due to the shallowness of the aquifer system, the surficial topography has a large effect on the flow and migration.
Figure 3.8 Boundary conditions across East Point peninsula (A-A')
Figure 3.9 Boundary conditions along East Point peninsula (B-B')
patterns" (Voss and Koch, 2001). Hodge’s report on Hornby Island showed that the water table has a similar profile to the topography of the islands, therefore, a coarse approximation of the water table, that follows the topographical profile is applicable.

A uniform recharge of $5.09 \times 10^{-9}$ m/s is applied at the top of the model, and it represents roughly 20% of the annual precipitation rate, which varies between 600 and 850 mm/year. Recharge is discussed further in the section “Groundwater recharge and spatial variations” of this chapter. A recharge flux boundary is defined by zero concentration, as no solute disperses or advects across this boundary. To determine the validity of using this particular boundary condition, a separate simulation was done; replacing the recharge boundary with atmospheric pressure = 0 (i.e., water table boundary). The water table was roughly delineated from previous studies (Suchy, 1998) and from data available in the BC Ministry of Water, Land and Air Protection Water Well Database. The results of this simulation are discussed further in Chapter 4.

A specified flux condition was also assigned for the bottom of the aquifer, as a no flow type of boundary.

As no data are available in terms of pumping rates in the aquifer, no internal sinks or sources of water were specified in the model.

Each of the previously-mentioned boundary conditions were located at natural locations where possible (seaside, topographical boundaries), or at a location consistent with the purpose of the model (bottom of the aquifer was chosen parallel to the flow lines and at a depth that does not affect the appearance of the interface). To eliminate instabilities or erroneous results, sharp changes in the boundary conditions were avoided. Boundary conditions for flow and solute transport simulations, for each cross-section are as shown in Figures 3.8, 3.9 and 3.10 for the two East Point cross-sections and for the entire island, respectively.

**Initial Conditions**

In order to run a transient model the initial conditions have to be first determined by running a steady-state simulation. Initial conditions for pressure, and concentration for solute transport and groundwater flow can not be specified separately, as they are a consequence of density and hydrostatic pressure variations. Initial pressures should be consistent with concentration distribution as well as with initial boundary conditions. One
way to determine initial conditions is to run a steady-state simulation using natural boundary conditions, which creates an initial set of pressure values. After repeated iterations, the proper configurations of pressure and concentration will be used as initial conditions for following transient simulations.

Mesh Discretization

An irregular mesh provided by SUTRA Pie was used for the purpose of discretization. It consists of irregular quadrilaterals that can have a variable density and variable dimensions at specified locations. Quadrilateral elements are defined by four nodes. The model domain or model outline was discretized by generating automatically within ARGUS One irregular meshes for each of the three cross-sections. The mesh was superimposed on the cross-sections for each area to be modeled. Various properties were assigned to each element: intrinsic permeability, the maximum and minimum longitudinal dispersivities, and the maximum and minimum transverse dispersivities.

Mesh size is governed, in part, by practical limitations and it is described by the Peclet number. Peclet number (Pe) is defined as:

\[
Pe = \frac{v \Delta x}{D_h} \tag{3.15}
\]

Where \( v \) = effective velocity of groundwater, \( \Delta x \) = dimension of the element, and \( D_h \) = hydrodynamic dispersion (Anderson and Woessner, 1992). For finite element methods, \( Pe \) is recommended to be \( Pe < 4 \).

Similarly, time discretization can be related to the Courant number, \( Co \), which is defined as:

\[
Co = \frac{v \Delta t}{\Delta x} \tag{3.16}
\]

Where \( v \) = effective velocity of groundwater, \( \Delta x \) = dimension of the element, and \( \Delta t \) = time step. Anderson and Woessner (1992) also recommended that time steps be specified so that \( \Delta t < \Delta x/V \) (or \( Co < 1.0 \)), which is equivalent to requiring that no solute be displaced by advection more than one grid cell or element during one time increment.
To minimize a variety of sources of numerical errors, the model grid should be designed using the finest mesh spacing and time steps possible, considering the previous conditions and given limitations on computer memory and computational time. To the extent possible, the grid should be aligned with the fabric of the rock and with the average direction of groundwater.

SUTRA determines automatically an optimized the bandwidth for each mesh, in order to optimize the efficiency of the solver. Mesh density varies as field conditions require; for example, a denser mesh is needed for zones with sharp changes in properties and a less dense mesh is needed for zones with relatively homogeneous properties. The number of elements and nodes differs from scenario to scenario and from model to model, each of which are described in more detail in the following chapter along with the results. A grid optimization study was performed; the results are included in Chapter 4.

**Input Parameters**

SUTRA requires as input:

1. The type of flow (saturated or unsaturated),

2. The type of the simulation (steady state or transient), and

3. A temporal discretization for transient simulations regarding the total length of the simulation, the maximum size of time steps within the simulation, the maximum number of iterations within a time step, and the iteration convergence criterion for the pressure and transport solutions.

The fluid and matrix parameters are defined in the following sections:

**Permeability**

Hydraulic conductivity (K) values were discussed in the previous chapter. However, SUTRA requires a permeability value (k) which is defined as:

\[ k = K \frac{\mu}{g \rho} \]  

(3.14)
Where: \( k \) = permeability (m\(^2\)), \( K \) = hydraulic conductivity (m/s), \( \mu \) = viscosity (kg/m\(\cdot\)s), \( \rho \) = density (kg/m\(^3\)) and \( g \) = gravitational acceleration (m/s\(^2\)).

As discussed previously, the geological formations on Saturna Island are considered heterogeneous in themselves, and an EPM concept will be used to assign hydraulic properties to each of them. As permeability values are abundant for some of the formations and scarce for others, the range of values used in the model will be varied according to the range of calculated values (roughly \( 10^{-10}\) m\(^2\) to \( 10^{-15}\) m\(^2\)) as well as in consideration of a range of values for similar geologic units according to the literature (up to \( k = 10^{-7}\) m\(^2\) t). Vertical permeability \( (k_z) \) is assumed to have the same value as horizontal permeability \( (k_x) \), and likewise, \( k_z \) is assumed the same as \( k_y \) (isotropic medium). As well, each formation is considered homogeneous.

**Dispersivity**

The dispersivity of a medium is characterized by two constants: longitudinal and transverse dispersivity. Conventionally, longitudinal dispersivity is a characteristic of the aquifer, which depends on the scale of investigation (0.01 - 1 x the scale of investigation).

**Porosity**

The transport equations employ effective porosity as one of the controlling factors of groundwater flow. No specific measurements were taken for Gulf Island formations to determine primary or secondary porosity. However, the literature mentions low porosity values, in the range of 5%, which is enhanced by fractures. Therefore, a value of 5% was used.

Other parameters are used in the model are summarized in Table 3.1.

**Groundwater recharge and spatial variation**

Saturna Island is a closed system, and groundwater recharge comes from precipitation. The main recharge zones are anticipated to be at higher elevations surrounding Mt. Warburton Pike and Mt. David, but geology, slope, and surface vegetation (arbutus, fir and oak trees, shrubs and abundant herbaceous vegetation) are expected to be important factors in determining the amount of recharge to the groundwater system. Major pathways for discharge to lower elevations and, ultimately,
<table>
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<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fluid compressibility ($\beta$)</td>
<td>$4.47 \times 10^{-10}$ ms$^2$/kg</td>
</tr>
<tr>
<td>Fluid base concentration</td>
<td>0 mg/l</td>
</tr>
<tr>
<td>Dissolved solids base concentration as a mass fraction</td>
<td>0 kg salt/kg of sea water</td>
</tr>
<tr>
<td>Dissolved solid base concentration of seawater as a mass fraction</td>
<td>0.0357 kg salt/kg of seawater</td>
</tr>
<tr>
<td>Density of freshwater ($\rho_l$)</td>
<td>1000 kg/m$^3$</td>
</tr>
<tr>
<td>Density of seawater ($\rho_s$)</td>
<td>1025 kg/m$^3$</td>
</tr>
<tr>
<td>Fluid viscosity ($\mu$)</td>
<td>0.001 kg/m s</td>
</tr>
<tr>
<td>Fluid diffusivity ($D_m$)</td>
<td>$1 \times 10^{-9}$ m$^2$/s</td>
</tr>
<tr>
<td>Coefficient of fluid density change with concentration</td>
<td>700 kg$^2$/kg</td>
</tr>
<tr>
<td>Cell thickness</td>
<td>1 m</td>
</tr>
<tr>
<td>Matrix compressibility ($\alpha$)</td>
<td>$10^{-6}$ Pa$^{-1}$</td>
</tr>
</tbody>
</table>

Table 3.1 Model parameters assumed for flow and solute-transport simulations

to the ocean are expected to coincide with fracture zones and faults because of their enhanced permeability. However, the exact role of geology and structure remains uncertain.

The amount of precipitation is highest between September and March, with amounts varying from 60 to 100 mm/month; while between April and September, it is much lower (26 to 50 mm/month). The mean annual precipitation rates, as estimated from historic data between 1983 and 1991, are shown in Figure 3.11. Data used to generate this graph were obtained from hardcopy records from the Saturna Island climate station. Average annual precipitation equals 666.87 mm/yr, and the negative slope of the trendline indicates a decrease in precipitation for this region over the period.
Figure 3.11 Variation of precipitation (bar graph) and temperature (line graph) on Saturna Island between 1981-1993.
of record, with an increase in temperature. Precipitation represents the only source of water into the groundwater system.

On Saturna Island, chemical data indicate that for East Point area pathways for groundwater tend toward the north edge of the peninsula as a consequence of layer dip (Allen and Suchy, 2001). The topography is relatively flat for the East Point region; it varies between 0 and 59 m. Recharge occurs in high areas, particularly to the east of the peninsula itself, and groundwater discharges towards the eastern tip of the island and out into the ocean primarily along the north edge of the island. At an island scale, groundwater flows from high areas, such as Warburton Pike, to lower areas (large valleys and the coast). With a Mediterranean climate, Saturna Island receives a low amount of precipitation, which can not assure significant quantities of freshwater as demand might require.

For the purposes of this study, it is assumed that the rate of recharge to the aquifer is spatially and temporally constant. The recharge rate is a function of precipitation and it is expressed as:

\[ RE = P - R - E \pm \Delta S \]  

(3.17)

where: \( RE \) = direct recharge, \( P \) = precipitation, \( R \) = surface runoff, \( E \) = evapotranspiration, and \( \Delta S \) = change in storage.

Because there have not been any studies for the Gulf Islands that have specifically attempted to estimate these various parameters, estimates of recharge are very rough. Current work is being done to estimate spatially-distributed recharge using HELP (Allen, Personal communication). For the purposes of this work, the sensitivity of the model to a range of recharge (10% to 50% of precipitation) will be determined.

**Surface water bodies**

With the exception of some swamps located at both high and low elevation areas, and Money Lake, which is a man-made lake that supplies water to residents in the Lyall Harbour area of Saturna Island, there are no other natural surface water bodies on the island. Ephemeral streams occur in some areas and flow in these streams is directly dependent on the amount of precipitation. In this study no surface water bodies are considered, and the aquifer is assumed to be recharged only by precipitation.
**Concentration Distribution**

A concentration profile for East Point peninsula was constructed (Figure 3.12) using data from Suchy (1998). The cross section is along strike of the Peninsula due to a higher amount of data available for this area. This cross-section captures the TDS variation with depth, showing a tendency of rapid increase in concentration with depth for area closer to the ocean boundary, and a smaller increase in TDS with depth at locations inland. The profile provides a rough image of the saltwater interface location, as well an image of the variation of salt concentrations with depth. The same variation with depth is observed for the entire island.

Plots of both chloride concentration and total dissolved solids versus depth show similar trends (Figures 3.13 and 3.14, respectively). Chloride concentrations vary between 8.3 mg/l to 2,800 mg/l (Suchy, 1998). Salinity (as TDS) increases with depth at depths below 100 m; TDS varies between 150 and 600 mg/L and increases with depth. At depths greater than 100 m, TDS concentrations are usually higher than 600 mg/L. Three zones can be observed:

- **Zone 1** is characterized by higher chloride concentrations (1990 mg/l) at 100-150 m depth, associated with direct salinization (through fractures). Zone 1 data are situated on the right side of the graphs (Figures 3.13 and 3.14).

- **Zone 2** is characterized by smaller chloride concentrations in between 50 and 200 m depth, which correspond to those groundwaters where cation exchange is dominant. Typically, these groundwater samples were collected inland. Zone 2 data are identified by a circle on the left hand side of the graphs.

- **Zone 3** is characterized by concentrations that increase linearly with depth due to the presence of the saltwater interface. The linear trendline equation will be used as calibration equation to determine the bulk position of saltwater wedge along the coast. Saltwater occurs at about 90 meters below surface (500 TDS).

The high TDS values and high Na and Cl concentrations on East Point region indicate that saltwater intrusion is a dominant. TDS variation with depth for the northern shore and the southern shore of East Point show some differences (Allen and Suchy, 1998). The northern shore data represent all three zones, and the trendline for zone 3 is similar to the one for all the chemical data (TDS values) from the entire island. On the
Figure 3.12 Salinity profile, accordingly to chloride concentrations measured on East Point area, Suchy, 2001.
Figure 3.13 Total Dissolved Solids variation with depth.

\[ y = 0.1402x + 12.57 \]

Figure 3.14 Chloride concentration variation with depth.

\[ y = 0.2058x + 44.836 \]
zone 3; all the other data are contained in the cation exchange zone (zone 2).
Nevertheless, the trendline obtained from these data results in a variation of TDS with
depth parallel to the one for the northern shore. The TDS variation with depth line for the
southern shore is situated above the one for the north shore, indicating higher total
dissolved solids concentration for this area. This is also supported by the difference in
the recharge amount that is lower on the southern side than on the north side of the
peninsula. The shape of the saltwater-freshwater interface obtained from numerical
simulations will be calibrated against a curve described by the trendline equation for
zone 3.

Summary

Based on the conceptual models presented in this chapter, steady-state and
transient simulations are run for the three cross-sections. Results, calibration and
sensitivity analyses are described in Chapters 4 and 5. As described in the previous
chapter, three different cross-sections for Saturna Island are used for modeling. Both
East Point cross-sections, across and along the peninsula, were run to determine
hydraulic parameters, and to calibrate them to the actual position of the interface.
Parameters established for East Point simulations are used previously in the model for
the entire island due to lack of numerous and accurate data for this cross-section.
CHAPTER 4 - STEADY-STATE SIMULATION RESULTS: CALIBRATION AND SENSITIVITY ANALYSIS

Steady-State Simulations

In this chapter, the effect of heterogeneity on the configuration of the saltwater-freshwater interface is investigated. The modeling process begins by using simple boundary conditions, a simple single layer geometry, and progresses to more complex models, involving multiple layer geometries, which describe field conditions more accurately.

Voss (1984) determined that saltwater intrusion is a non-linear problem and that steady-state conditions can only be reached through time steps. Therefore, the steady-state conditions of a system are determined by long-term transient simulations from arbitrary initial conditions, until the system stabilizes (Souza and Voss, 1987). In this thesis, simulations of saltwater intrusion begin by running steady-state flow and steady-state transport. These preliminary simulations are done in order to determine pressure and velocity distributions that can be used as the initial conditions for a transient simulation. Thus, at time zero, seawater starts to intrude the system across the sea boundary, and the "true" steady-state position is reached by progressive intrusion through time. The model is run until the system stabilizes (i.e., until there is no significant change in concentration contours with time). The equilibrium position indicates that both the flow and the transport mechanisms have reached the steady-state (Voss, 1984).

Steady-state simulations are run to:

- Determine the appropriate parameters for future simulations (k values, recharge amount, time steps, diffusivity, and mesh discretization dimensions).
- Determine appropriate initial pressure and concentration distributions for transient simulations.
- Determine the effect of heterogeneities on the shape of interface
Reproduce the real bulk salinity profiles, instead of the exact position of the interface, which is under the effect of pumping and tides.

**East Point Area Simulations**

Steady-state simulations were undertaken first for the East Point area of Saturna Island. This area was chosen because most of the chemistry data and water level measurements are available for use in calibrating the model. As discussed in the previous chapter, two 2-D cross-sections models for East point area were constructed. The first profile (A-A' profile as shown in Figure 3.5) extends across the peninsula, and is 1500 m long and 300 m deep. This profile is referred to as "Across the Peninsula". The second profile (B-B' in Figure 3.6) extends along East Point peninsula, and is 1500 m long and 300 m deep. This profile is referred to as "Along Strike of the Peninsula". Results for each profile are discussed separately.

**Across the Peninsula**

Three sets of simulations were undertaken across East Point peninsula in order to investigate the effect of heterogeneity: homogeneous, heterogeneous two layers, and heterogeneous three layers. The layer geometries are shown in Figure 4.1 The single layer domain is used to simulate the effects of variation in the magnitude of permeability; testing the range of values obtained from the pumping test analyses. The heterogeneous two layer model is used to investigate the effect of adding a second layer to the domain, while the three layer model is used to investigate the addition of a high permeability interlayer. Modifications to the model parameters are described as different scenarios within each section. A sensitivity analysis was also undertaken to test the effect of variations in dispersivity and recharge, as well as the type of boundary condition applied to the top surface of the model (i.e., specified flux or specified pressure). Results of these simulations are described separately.

Simulations for East Point started by using simple parameters and boundary conditions; complexity was added afterwards. A finite element mesh with an element
Figure 4.1 East Point across peninsula models, scenarios that includes different types of representation of geology: homogeneous, 2 layers and 3 layers models.

Distance (m)
dimension of 20 (i.e., 20m x 20m) was used initially; mesh density was increased to 10m x 10m both at the sea boundaries and in specific areas where saltwater intrusion is expected to occur. The non-uniform grid is comprised of 3888 nodes and 3742 elements (Figure 4.2). This particular mesh density did not result in numerical instability; therefore it was used for all East Point simulations. All simulations were carried out for isotropic conditions, the ratio of $k_x/k_y = 1$. Dispersivities were assigned the following starting values (Voss, 2003 personal communications), and were subsequently varied during the sensitivity analysis:

- Longitudinal dispersivity = 10 m
- Transverse dispersivity = 1 m

The convergence criterion for the pressure boundary condition factor equals $10^{-5}$ for all simulations.

**Homogeneous Single Layer Model**

In the homogeneous single layer model, the domain is considered to be comprised of a single geologic formation. Simulations were carried out for permeability in the range of $2.2 \times 10^{-8}$ m$^2$ (maximum value corresponding to an increase of the average calculated permeability by 6 orders of magnitude) to $2.2 \times 10^{-14}$ m$^2$ (average k obtained from Theis method); all values were based on the results of aquifer tests conducted in the Geoffrey Formation, as described in chapter 2. The reason for using this particular range of values is the fact that the geology of East Point is comprised of the Geoffrey and Northumberland Formations. However, no aquifer tests were conducted in the Northumberland Formation, which is anticipated to have a higher k than the Geoffrey based on the fracture study.

The primary objective of these models is to determine what effect permeability has on the configuration of the freshwater-saltwater interface, and to establish whether or not the hydraulic parameters derived directly from pumping test analyses can be used in a model and yield reasonable results. This is important because the conceptual model is based on the EPM approach, while the actual system is fractured.
Figure 4.2 Finite element meshes for East Point is represented by non-uniform grid comprised of 3888 nodes and 3742 elements.

Figure 4.3. Velocity distribution at steady-state (scenario 1). Low values of velocity at greater depth, and higher values at the top where recharge occurs, and at the sea boundaries where saltwater intrudes the island.
Variations in dispersivity and recharge are also considered. For the various scenarios, recharge is held constant, and represents 20% of the precipitation amount. Only the most significant results are presented herein; the ones showing the effect on saltwater intrusion of varying permeability, dispersivity and recharge.

**Scenario 1- low permeability**

In the first scenario, permeability equals the average value obtained for Geoffrey Formation using the Theis method of analysis \( k = 2.2 \times 10^{14} \text{ m}^2 \). The velocity distribution (Figure 4.3) corresponds to the expected type of flow, which is controlled by the topography of the area. The main recharge area is located at the highest elevations on East Point; primarily on the southern side where topography attains an elevation of 60 m.a.s.l. Groundwater flows from these highest points and laterally out towards the ocean boundaries. At depth, velocities are parallel with the bottom of the model, showing that the chosen depth does not influence the flow pattern. Velocity varies between \(2 \times 10^{-12} \text{ m/s}\) to and \(5 \times 10^{-7} \text{ m/s}\) as a result of pressure and recharge values. The higher velocities occur on the southern side of the peninsula, where topography is steeper and where the greatest change in pressure with depth occurs. At steady state, lines of equal pressure (isobars) are parallel and horizontal, and increase with depth as boundary conditions specify.

To determine the steady-state position of the interface, a steady-state groundwater flow and transient transport simulation was run for a period of 1,000 years using a time step of one day. Steady-state was reached after 700 years as evidenced by the cessation of movement of the interface from the sides. The configuration of the interface is shown for year 1,000 (Figure 4.4). The results show that saltwater intrudes the island only to a limited extent along the sides of the model domain. This is not consistent with what we observe today, based on measured concentration values. Figure 4.5 shows a graph of the theoretical TDS concentration depth profile (from the actual data as described in the previous chapter) versus the model results. Data are shown as a function of distance from the north (right) coast of the model domain. A comparison between the TDS correlation curve and TDS vs. depth obtained for this scenario shows a poor correlation. The results of this simulation are situated below the
Figure 4.4 Concentration distribution, scenario 1. Low permeability – Theis method, recharge 20%.
Comparison between resulted and observed TDS concentrations for Scenario 1

Figure 4.5. Comparison between TDS calibration curve and TDS variation with depth for scenario 1 for different distances from the coast.
calibration curve. However, Figure 4.5 shows most of the data above the curve due to the small values of simulated TDS.

The results indicate that the low values of permeability, which were determined directly from the Theis analysis, do not allow the saltwater to intrude the island; the interface is positioned only on the sides. Simulations were run for extended times (10,000 years) and recharge was lowered in order obtain a more extensive intrusion. However, at these low values of permeability, the model does not appear to accurately represent field conditions. One possible explanation for the small magnitude of intrusion may be a result of the fact that the movement of solute is mainly controlled by dispersion and advection. If permeability is too low, then the velocities are low, and both advective and dispersive transport are minimized. In the case of small permeability values, solute transport is dominated by molecular diffusion rather than mechanical dispersion. Molecular diffusion was set at a realistic value of $1 \times 10^{-9}$ m$^2$/s. Longer simulation times with a finer mesh may result in greater diffusive transport; however, for the present case convergence to steady-state was observed.

**Scenario 2 – high permeability**

The results of the previous simulations indicated the necessity of using permeability values up to few orders of magnitude greater. A series of simulations were undertaken using increased values of $k$; only the best results are described here. Simulations were undertaken using a high permeability of $2.2 \times 10^8$ m, which reproduces approximately the present day position of the interface (Figure 4.6). The average permeability value obtained from Theis method was increased until a good match (same order of magnitude for TDS at different depths) was observed between field measured TDS values and calculated TDS values.

The best match was observed when $k$ was increased to up to 6 orders of magnitude higher than the average $k$ value of $2.2 \times 10^{14}$ m. However, this permeability corresponds to an increase of only 2 orders of magnitude relative to the maximum calculated $k$ value of $1.7 \times 10^{-10}$ m (Gringarten method). As discussed in Chapter 2, the Gringarten method yields slightly higher values of $K$ because it is a linear flow model, and as such, takes into account the presence of any major fractures.
Figure 4.6 TDS concentrations after 1000 years using maximum values for permeability and a recharge of 20% from precipitation amount- Scenario 2.
The steady state position is reached after 700 years of simulation. Results are shown for year 1,000 (Figure 4.6). The maximum depth of the interface is simulated to occur near the centre of the peninsula, where it attains a depth of -100 m, and is consistent with the deeper presence of circulating fresh groundwater, relative to the ocean edges. Field data support the fact that this particular simulation is more representative of actual conditions. Using the TDS calibration curve, saltwater is expected to occur at about 95 m below sea level. Figure 4.7 shows a comparison between the expected TDS concentration and the model results. The best match occurs for the observation wells situated 200 m to 600 m away from the margins of the model (near the actual shore of the island). Bigger differences occur at the centre of the island where simulated concentrations are lower than actual concentrations. However, because there are few actual data from the central portion of the peninsula, and the island itself, the calibration curve may be somewhat biased in that it may best represent the higher concentrations present nearer to the coast. Moreover, the best match is observed within the first 50 m below the topographical surface, which indicates that this k value is in the right range, since most of the chemical data are collected at depth between 0 and 50 m below m.a.s.l.

Another factor that may influence the results is the amount of recharge. In the simulations the average annual precipitation was used, but the actual chemical data were collected during the summer, when recharge is lowest. The effect of this discrepancy would be to simulate perhaps lower overall concentrations. Notwithstanding, the higher value of permeability appears to give more realistic results for the position of the interface.

To summarize, the best match between calculated and observed concentrations occurs for a permeability of $2.2 \times 10^{-8}$ m$^2$.(corresponding to an increase of 2 orders of magnitude higher than maximum k obtained from Gringarten method) The highest values tend to overestimate somewhat the intrusion of saltwater, but nevertheless provide better results that those using the lower k values. Obtaining the best match for the higher range of values of permeability indicates that fractures have a major role in the amount of intrusion. In contrast, low range permeability values appear to limit the amount of intrusion, perhaps because they reflect more the contribution of porous media flow rather than fracture flow.
Figure 4.7. Comparison between TDS calibration curve and TDS variation with depth for scenario 2. The best match occurs close to the northern shore and for the first 50 m from the surface.
Scenario 3 – Recharge

Two simulations were conducted to test the effect varying the magnitude of recharge and the accuracy of applying a recharge boundary to the top portion of the model. In both cases the permeability of the single layer is \( k = 2 \times 10^{-8} \text{ m}^2 \).

Recharge was increased to 50% of total precipitation. Only small differences in the position and magnitude of intrusion are observed (Figure 4.8). At a distance of 100 m from the shore TDS concentrations are lower compared to those obtained at a recharge of 20%. These results are anticipated given the greater inflow of freshwater into the system. Slightly larger differences are noted moving progressively inland, towards the centre of the peninsula. Here the effect of recharge might be expected to have a greater effect on the position of the interface because of the larger potential volume of freshwater in that portion of the aquifer. A decrease of about 7% in TDS concentrations is observed with higher recharge.

In a separate simulation, a specified atmospheric pressure boundary condition (\( P = 0 \text{ atm.} \)) was assigned to the top of the model domain, in the place of a flux (recharge) boundary condition. All other boundary conditions and model parameters were kept the same as in the previous simulations. The specified pressure was assigned to the position corresponding to water table. The water table was defined by converting the measured heads (Suchy, 1998) to equivalent freshwater heads as described below.

Usually fluid pressure in an aquifer determines the hydraulic head, which itself is measured as the column of freshwater in a piezometer. Aquifers in coastal zones are characterized by a non uniform density distribution, thus, measured heads vary according to the salinity of the water in the well. Therefore, piezometric heads for saltwater must first be converted to equivalent freshwater heads. This conversion is necessary because salinity lowers the elevation to which groundwater will rise in a water well. The following equation is used to convert to equivalent freshwater head:

\[
\varphi_f = z + h_f, \text{ and: } h_f = \frac{\rho^*}{\rho}h
\]  \hspace{1cm} (4.1)

where

\( \varphi_f \) = freshwater head of the observation well (m),
Figure 4.8. Comparison between TDS calibration curve and TDS variation with depth for scenario 3. Results show a small difference in concentration with depth between simulations using a different percentage of precipitation as recharge (20 and 50%).
\[ z = \text{elevation head (m)}, \]
\[ h_f = \text{pressure head (m)}, \]
\[ \rho = \text{density of saltwater (g/cm}^3)\]
\[ \rho_i = \text{density of fresh water (g/cm}^3)\]
\[ g = \text{acceleration of gravity (m/s}^2)\]

This simulation was done in order to test the validity of recharge flux. The velocity profile shows a flow controlled by the topography (Figure 4.9). Values calculated for velocity, using atmospheric pressure as boundary condition, are in the same range as the ones calculated using a recharge boundary, and fall in the range \[ v = 4.00 \times 10^{-3} \text{ to } 2.51 \times 10^{-6} \text{ m/s}. \] Steady-state is similarly reached after 700 years, therefore, saltwater intrudes the aquifer at approximately the same rate. Figure 4.10 shows the concentration profile at 1000 years, and is essentially the same as the previous simulation. TDS values are equal to those determined using a recharge of 20% of total precipitation. These results increase confidence in the recharge boundary, and suggest that either a specified flux or a specified pressure boundary can be used. For consistency, a specified flux boundary with a concentration of 0 mg/L TDS is used in all the following simulations.

**Scenario 4 - Dispersivity**

The exact values for longitudinal and transverse dispersivity are difficult to determine, because these model parameters are a function of scale, transient effects, and fingering (Anderson and Woessner, 1992). In order to investigate the effect of changing dispersivity on the character of the intrusion, dispersivity was varied between 10 and 400 in longitudinal direction \((\alpha_L)\), and between 1 to 100 in transverse direction \((\alpha_T)\), keeping the ratio of changes to 1/10.

Higher dispersivity values \(\alpha_L = 100\) and \(\alpha_T = 10\) result in a more extensive interface, and a higher rate of intrusion of saltwater into the aquifer (Figure 4.11). The bigger differences in calculated TDS occur near shore, while moving inland the differences are smaller. However, confidence in the calibration curve decreases.
Figure 4.9. Velocity profile at steady-state (scenario 3), using atmospheric pressure at water table, shows the same values as using as boundary condition the recharge. Velocities are in the same range in both cases increasing the confidence in the purposed model.

Figure 4.10 Concentration distribution, scenario 3. Maximum permeability – Gringarten method, using as boundary condition atmospheric pressure at water table
Figure 4.11 Concentration distribution, scenario 4. Maximum permeability – Gringarten method, higher dispersivity values ($\alpha_L = 100\text{m}$ and $\alpha_T = 10\text{m}$)
towards the centre of the island as discussed earlier. The greater amount of mixing of saltwater and freshwater, which would occur with higher dispersivity values, results in a larger interface than in the previous cases (lower $\alpha_l$ and $\alpha_T$). Longitudinal dispersivity has the most significant effect on the length of the transition zone. A decrease of $\alpha_l$ results in a greater TDS gradient; TDS increases more rapidly with depth using a smaller dispersivity, and the best match is observed for $\alpha_l = 10$ and $\alpha_T = 1$, values that will be used for further simulations.

**Heterogeneous Two Layer Model**

In this simulation, a two layer geometry is considered. The top layer of the model corresponds to the average permeability for Geoffrey Formation obtained from Theis method of analysis. The second, or bottom layer, was assigned a one order of magnitude higher permeability that would correspond to a highly fractured mudstone layer (Northumberland Formation). The dip of both layers is 20 degrees. Layer one has a thickness of 160 m, and layer 2 has a thickness of 200 m (as shown in Figure 4.1).

The permeabilities used in the model are as follows:

- $k_{\text{Geoffrey}} = 2 \times 10^{-14}$ m$^2$
- $k_{\text{Northumberland}} = 2 \times 10^{-13}$ m$^2$

Recharge represents 20% of precipitation amount.

In general, the velocity distribution (not shown) is similar to that of the homogeneous models, indicating that the shape of the interface is largely controlled by topography. However, there are subtle differences in the magnitude of the velocities and the appearance of the distribution, particularly at the model edges and at the contact between the two layers. Inserting the second layer increases groundwater velocity for the entire model to up to 2%; higher values are observed in the more permeable layer. The vector profile approximates the position of saltwater-freshwater interface at the sides of the island and also shows the groundwater divides as being controlled by the two higher elevation points that occur near the centre of the peninsula.
The chloride concentration contours indicate that the range of permeabilities used are too small to allow saltwater to intrude the island. However, an increase in the permeability of the second layer results in a larger amount of intrusion than using a single layer (Figure 4.12). An increase in k for the second layer corresponds to a greater ease of entry of the saltwater into the freshwater aquifer, nevertheless the TDS values can not be compared since the resulted TDS concentrations are in the range $10^{-20}$ to $10^{-74}$ kg/l.

Therefore using a lower permeabilities range of values the results are consistent to the homogeneous model results that show a smaller amount of intrusion than the expected ones.

**Heterogeneous Three Layer Model**

In the last chapter, a conceptual model for East Point was discussed. The geology of East Point is represented by two formations - Geoffrey and Northumberland. The contact between the two layers is represented by a transition zone, consisting of interlayered sandstone and mudstone. As shown Figure 3.7, this transition zone is characterized by a permeability that is higher than the sandstone dominant formation above (Geoffrey Formation) and the mudstone dominant formation below (Northumberland Formation). Within the model, this transition zone is represented as a separate layer, thus a model domain consisting of three layers was considered.

**Scenario 1**

Numerous simulations were carried to determine which k values reproduce the actual position of the interface.

Considering the same range of k values ($2 \times 10^{-14}$ to $6 \times 10^{-13}$ m) as used previously a model was run to determine the steady-state position of the interface. As shown in Figure 4.13 this scenario can not reproduce the actual situation and results in an underestimation of the saltwater movement inland. Based on previous findings better results are expected to be obtained for a higher range of permeability. Thus, for this scenario the average k (Theis method) values were
Figure 4.12 Concentration distribution, 2 layers scenario 1. Maximum permeability fractured sandstone formation, recharge 20% of precipitation amount.

Figure 4.13 Velocity profile using interlayers characterized by a higher permeability, shows an increase of velocities in this area.
increased to up to 5 orders of magnitude for sandstone-dominant formation, and subsequently, to an order of magnitude higher for mudstone dominant formations, whereas the interlayers were assigned the highest k as follows:

\[ k_{\text{Geoffrey}} = 2 \times 10^{-9} \text{ m}^2 \]
\[ k_{\text{Interlayer}} = 7 \times 10^{-8} \text{ m}^2 \]
\[ k_{\text{Northumberland}} = 2 \times 10^{-9} \text{ m}^2 \]

Steady-state is obtained after 500 years. Saltwater intrudes the island to a greater extent as shown in Figure 4.14.

Simulated TDS values are of the same order of magnitude as the observed TDS values (Figure 4.15); however, depending on the distance from the shore, the TDS variation with depth is either under or over-estimated. This behaviour is expected to occur since the data used to calibrate the results come only from areas along the shore.

The 500 mg/L isochlor is situated at a depth of approximately 30 m below the water table (Figure 4.14). The chemical transition (mixing) zone is larger than observed previously due to the third layer having a higher permeability. This is a consequence of the fact that water is able to move more quickly and mix more readily with seawater.

The best representation of the hydrogeology, as evidenced by the best fit between the observed concentrations and the model concentrations (Figure 4.15) corresponds to the case where layering is introduced. However, to determine the best match between present situation and simulated position of the interface calculated permeability values should be increased to up few orders of magnitude.

This adjustment of the permeabilities could be a direct consequence of two factors:

- Underestimation of k values that were directly derived from transmissivity results. Hydraulic conductivity was calculated as a ratio between T and the length of the open borehole, whereas in fractured rock groundwater moves preferentially through fractures (which summed result in a considerably smaller thickness than the open borehole).
- Calculation of hydraulic conductivity using the radial flow, rather than linear flow that describe the flow through fractures.
Total Dissolved Solids Distribution after 1000 years

Figure 4.14 Concentration distribution, 3 layers model. Recharge represented as 20% of annual precipitation.
Figure 4.15 Comparison between TDS calibration curve and TDS variation with depth for 3 layer model.
The model that reproduces the present position of the saltwater-freshwater interface under Saturna Island is characterized by a three layer model having a lower permeability sandstone dominant formation over a higher permeability mudstone dominant formation, with a very high permeability interbedded layer. Based on a coarse calibration, permeability values to be use for the cross section for the entire island will be in the range of $10^{-8} \text{ m}^2$ to $10^{-9} \text{ m}^2$.

**Along Strike of the Peninsula**

Three sets of simulations were undertaken to model flow and transport along a cross-section that extends along strike of East Point peninsula. These simulations were done in order to provide a pseudo 3-D view of saltwater intrusion beneath the peninsula. Cross-section B-B' is located as shown in Figure 3.4, and sets of homogeneous, heterogeneous (two and three layers) models were simulated; however, only the three layer results are shown in this section. The layer geometries are shown in Figure 4.1.

The simulations were started by using simple parameters and boundary conditions. A finite element mesh with element dimensions (20m x 20m) was used. The non-uniform grid was comprised of 3447 nodes and 3280 elements (Figure 4.16). All simulations were carried out for isotropic conditions, the ratio of $k_x/k_y = 1$. Dispersivities were assigned as follows:

- Longitudinal dispersivity = 10m
- Transverse dispersivity = 1m

The convergence criterion was set to $10^{-5}$ (units) for all simulations. Only the relevant simulation results are discussed.

**Three Layers**

For this simulation the following permeabilities were used:

- $k_{Geoffrey} = 2 \times 10^{-9} \text{ m}^2$
- $k_{interlayer} = 7 \times 10^{-8} \text{ m}^2$
- $k_{Northumberland} = 2 \times 10^{-8} \text{ m}^2$
Figure 4.16 Regular mesh represented by elements with a size of 20x 20m.

Figure 4.17 Velocity distribution at steady-state (scenario 1), shows that the highest velocities occur at the ocean boundaries, which are represented by specified pressure boundary conditions.
Figure 4.17 shows the velocity profile; higher velocities occur at the ocean boundaries, whereas inland the groundwater velocity decreases. TDS concentration results for a 3 layer model are depicted in Figure 4.18, whereby each layer is characterized by the same k values as were used for A-A' model (3 layers). For the previous model, these were the best k values that reproduced the position of the interface. The simulated TDS values are similar at different distances from the coast, and the appearance of the TDS variation with depth curve (Figure 4.19) indicates a small increase in TDS for the first 50 m below sea level. An increase in saltwater concentration is observed below 50 m, followed by an abrupt and linear increase of TDS. Furthermore, between 0 and -50 m below m.a.s.l TDS values fall in the same range as the chemical analyses indicated, and in deeper areas a greater amount of saltwater occurs. This scenario reproduces only partially the shape and position of the interface; however, it indicates that higher k values are more suitable to represent the geology of Saturna Island in order to obtain realistic modelling results.

Nevertheless, the simulation results for both the across the peninsula and along strike of the peninsula demonstrate that the primary direction of intrusion is south–north (A-A'), rather than east-west (B-B'). East Point peninsula has relatively small dimensions (10 km N-S and 2.5 km E-W). Therefore, the exact mixing behaviour is difficult to capture given the complexity of layering present. The Nanaimo Group formations dip at 20° to the North, or perpendicular to the orientation of the B-B' cross-section. Therefore, reconstructing a 3-D view of saltwater mixing from two separate and perpendicular cross section simulations does not adequately reproduce the entire amount of intrusion. Scenarios that represent the interlayer as a higher permeability medium similarly do not show the important differences.

**Saturna Island profile C-C”**

**Effect of Permeability**

A heterogeneous model was run to determine the position of saltwater-freshwater interface beneath the entire island. This simulation used different permeabilities to represent the fractured sandstone and mudstone units. Each layer was assumed to be isotropic, $k_x/k_y = 1$. The geology was represented as a succession of
Figure 4.18 TDS concentrations at steady-state (scenario 1), after 650 years, TDS= 500 mg/l occurs at various depths as a function of topography and permeability variations.
Figure 4.19 Comparison between TDS calibration curve and TDS variation with depth for B-B' cross-section along the strike of the peninsula, showing smaller TDS concentrations further inland.
layers (excluding the interlayer to avoid numerical instabilities due to numerous changes in properties). In this section, two sets of simulations are discussed. The first simulation uses the maximum k values in the range of those obtained from the pumping test data (Scenario 1).

The second simulation uses k values that are 5 orders of magnitude higher than the average calculated k values (Scenario 2). Steady-state simulations were run to reproduce the pressure and velocity distribution that provide the initial conditions for transient simulations. The irregular mesh consists of 3559 nodes and 3393 elements as illustrated in Figure 4.20. Data from 5 wells provide calibration data, along with the TDS trend line estimated from the East Point data.

**Scenario 1 (low k values)**

The hydraulic conductivities from pumping test data for Nanaimo Group Formations were used for this simulation. Where a particular formation is not represented, an appropriate k value was assigned to that formation, based on whether the formation is sandstone or mudstone dominant. The values are as follows:

\[
\begin{align*}
  k_{\text{Sandstone}} &= 2 \times 10^{-14} \text{ m}^2 \\
  k_{\text{Mudstone}} &= 2 \times 10^{-13} \text{ m}^2
\end{align*}
\]

The velocity profile shows larger magnitude vectors along the two coasts, as might be predicted based on the topography (Figure 4.21). Saltwater intrudes the island preferentially in the areas where the mudstone formations coincide to the pressure boundary condition (corresponding to higher permeabilities) as shown in Figure 4.22. However, as was observed in the previous simulations, the amount of saltwater intrusion is underestimated using a low range of k.

A comparison of the observed TDS concentrations and the model simulated concentrations supports the conclusion that the low k values do not provide a good representation of the interface position (Figure 4.23). Simulation results show that the interface is situated below its actual position. Note that in the figure only the higher range of TDS in captured, since the other TDS concentrations were too small to be plotted \((10^{-49} - 10^{-10} \text{ kg/l})\).
Mesh for Across Saturna Island model (C-C')

Figure 4.20 Irregular mesh containing 3559 nodes and 3393 quadrilateral elements

Velocity profile at steady-state

Figure 4.21 Velocity profile at steady-state (scenario 1), showing low values of velocity at greater depths, and higher values at the top – recharge boundary, and at the sea boundaries where saltwater intrudes the island.
Figure 4.22 TDS concentrations at steady-state (scenario 1), after 1000 years. TDS = 500 mg/l occurs at depths varying as a function of topography and permeability.
Figure 4.23. Comparison between calculated and simulated TDS concentrations across the entire island, scenario 1
Scenario 2 (high k values)

Based on the previous results, a second scenario was run to examine the effect of increasing k for each formation by 4 orders of magnitude. Permeability values are as follows:

- $k_{\text{Sandstone}} = 2 \times 10^{-10} \text{ m}^2$
- $k_{\text{Mudstone}} = 2 \times 10^{-9} \text{ m}^2$

Velocities vary in different regions, based on the thickness and permeability of each unit or the boundary conditions. Higher velocities occur near the sea boundaries, and within layers characterized by higher permeabilities. Preferential paths occur within the Northumberland, Spray and de Courcy Formations, and create upward flow in the central part of the island. Here velocities range from $10^{-6}$ to $10^{-8}$ m/day in horizontal direction, and $10^{-7}$ to $10^{-11}$ m/day in vertical direction. TDS concentrations of 500 mg/L (Figure 4.24) occur at depths between 100 and 170 m. The best match is observed on the northern side of the island, where the geology consists of the Northumberland, Geoffrey and Spray Formations; the ones used in the previous simulations. A comparison of the curves shows a similarity between observed TDS and simulated values at greater depths; about 150-200 m below ground surface, as illustrated in Figure 4.25.

A good match was observed for scenario 2 suggesting that the k values may be more representative of the aquifer. Nevertheless even higher k values are expected to result in more accurate results in terms of TDS concentration variation with depth. Therefore, another simulation was run using a higher range of permeabilities (Scenario 3).

Scenario 3

Scenario 3 was run to examine the effect of increasing k by up to 5 orders of magnitude relative to the values determined from Theis (one order of magnitude higher than scenario 2). Permeability values are as follows:

- $k_{\text{Sandstone}} = 2 \times 10^{-9} \text{ m}^2$
- $k_{\text{Mudstone}} = 2 \times 10^{-8} \text{ m}^2$
Figure 4.24. TDS concentrations at steady-state (scenario 2), after 1000 years.
Figure 4.25 Comparison between calculated and simulated TDS concentrations across the entire island, scenario 2.
The 500 mg/l isochlor occurs at depths between 180 and 250 meters (Figure 4.26) below m.a.s.l. as a consequence of topography (water table profile) and layering (permeability variability). Compared to the previous simulation, a better match is observed in between simulated and observed TDS values below 50 m.a.s.l., but TDS values are higher than the chemistry data would indicate at shallower depths (Figure 4.27).

Comparing all three scenarios the best match is observed when the higher permeability values are used. Both scenarios 2 and 3 give reasonable results, and therefore, the values for k for these models will constrain the values used in the transient simulations. The discrepancies might be anticipated because the calibration curve was constructed using mostly data from the East Point area; only a few samples were collected and analyzed from the center of the island. Therefore, simulations results for the entire island can not be constrained as well as the East Point zone using the calibration curve.

Sensitivity Analysis

A sensitivity analysis was performed to determine the response of the model to changes in permeability, different types of layering, and changes in recharge, dispersivity, and matrix compressibility.

Prior both effects of layering and recharge variations were describe as part of different scenarios. It was found that recharge does not play a major role in the appearance of saltwater-freshwater interface, while different scenarios of layering plays an important role in both dimension and magnitude of intrusion.

Permeability

Permeability values were constrained by the values obtained from the pumping tests analyses performed throughout the Gulf Islands. Results show that permeability represents one of the controlling factors on the depth and configuration of the interface. Where permeability is low, saltwater does not intrude the island to the same extent as when higher values are used.
Figure 4.26. TDS concentrations at steady-state (scenario 3), after 1000 years.
Comparison between calibration and resulted TDS curves - Scenario 3

Figure 4.27. Comparison between calculated and simulated TDS concentrations across the entire island, scenario 3.
Figure 4.28 shows the range of hydraulic permeability values calculated using the different methods of analysis, and the k values (designated by an X) that result in the best match between the simulated TDS values and the calibration curve for Saturna Island. In some cases, k values were higher for sandstone dominant formations than for the mudstone dominant ones. Generally, mudstone dominant formations have permeabilities in the order of $10^{-9}$ to $10^{-11}$ m$^2$, while sandstone dominant formations are represented as permeable formations in the range of $10^{-10}$ to $10^{-11}$ m$^2$.

The simulation results suggest that high k values are more realistic, and that k is an important factor controlling the magnitude of the intrusion. Permeability values few orders (up to 5 orders of magnitude) of magnitude higher than the average calculated hydraulic conductivity for Nanaimo Group Formations provide the best results, as evidenced by the good correlation between model-derived TDS and observed TDS (differences in TDS up to 20%). Permeability values determined directly from the pumping tests, and which represent an equivalent porous media model do not appear to adequate take into consideration the effect of fracturing, and consequently, do not reproduce the actual state of saltwater intrusion. By increasing the values of k, a more realistic representation of the interface position is obtained.

**Dispersivity**

Anderson (1979) noted that at a small scale dispersivities are in the range of few cm to m, while at regional scales dispersivities are in the range of hundreds of meters. To investigate the influence of dispersivity on the concentration distribution, higher values were used (up to 200m). Steady-state was obtained after 650 years using the higher dispersivity values. Higher dispersivity values result in a more extensive interface in both vertical and horizontal directions. Higher concentrations occur at the sides of the island, where the aquifer is in direct contact with saltwater. Similar results were obtained in the East Point simulations; an increase in transverse dispersivity (from 10 to 200) resulted in a larger interface, and produced an increase in TDS from 300 to 560 mg/l (observed locations —wells EP-27 and EP - 30) (same increase in both wells). So, $\alpha_z = 10$ and $\alpha_T = 1$ represent the best solution to reproduce the trends in TDS variation with depth.
Figure 4.28. Limits of calculated hydraulic conductivity values, and used values for numerical modeling.
Other Parameters

Ranges in other parameters, such as porosity, solid matrix compressibility, or fluid viscosity are constrained by literature reported values, and therefore, a sensitivity analysis could not be undertaken to evaluate the effect of varying these parameters.

In the following chapter, two transient simulations will be undertaken in order to model the evolution of the freshwater-saltwater interface through recent geologic time. The first two scenarios used to simulate saltwater intrusion across the island will be used as starting point to test the Pleistocene history theory. The parameters used in the simulation are determined by the best set of parameters from these steady-state simulations. It is important to note however that calibration data were few in number, and there are a large number of parameters to be defined. Consequently, some degree of uncertainty will accompany the results of the transient simulations.
CHAPTER 5- TRANSIENT SIMULATION – PLEISTOCENE EVOLUTION

As stated in previous chapters, the position and shape of saltwater–freshwater interface is a result of many factors. One of the most significant factors is climate change, a factor that determines variations of sea level and of the amount of recharge. With regard to sea level, a sea level rise would generally result in an advance of saltwater inland, while a drop of sea level enables freshwater to push the salty water towards the ocean, increasing the size of the freshwater lens.

During the late Pleistocene, the Gulf Islands were partially submerged for a period of about 500 to 1000 years (evidence in support of this time frame is provided later). From a hydrogeological perspective, this hypothesis is supported by the groundwater chemistry (Allen and Suchy, 2001) and stable isotope data ($^{18}$O and $^{34}$S in dissolved sulphate, and $^{18}$O and $^2$H in water) (Allen, in Press). The chemistry shows that cation exchange is a dominant chemical mechanism on the islands, whereby Ca-HCO$_3$ rich freshwater undergoes Na exchange. This Na-enriched water is present in groundwater throughout most of the island. At depth and near the coast, there are higher concentrations of Cl, consistent with the presence of a saltwater wedge. Na is speculated to originate from remnant seawater, which was introduced (along with Cl) into the rock sequence during the Pleistocene when the islands were submerged (Allen and Suchy, 2001). Isostatic rebound, following submergence, has progressively removed Cl, by flushing the aquifer with fresh (HCO$_3$ rich) infiltrating groundwater. It is speculated that Na largely remained in place because of sorption to clay minerals, but is now being released as fresh water continues to infiltrate.

In order to demonstrate the validity of this mechanism it is necessary to assess the time frame over which the sea level varied during the Pleistocene, and to determine the concentration profile after 500 -1000 years of submergence followed by a rapid re-emergence of the islands.

In this chapter, two models are developed to investigate the paleohydrogeological evolution of the saltwater-freshwater interface, and hence, saline groundwater distribution on Saturna Island.
Based on the conceptual model, uncertainties are expected to derive from the following assumptions:

- Saturna Island was submerged below sea level to an elevation of 150 m above the present day level for a period of 500 years (a hypothesis that is not constraint by rigorous data).

- Recharge is constant during the last 12,000 years. This is unlikely, but nevertheless considered to be of less significance because a sensitivity analysis (Chapter 4) showed that the amount of precipitation does not determine major changes in the position or shape of the saltwater intrusion.

Using both conceptual models simulated in the previous chapter (Scenarios 1 and 2), transient simulations are run to test the hypothesis “the current distribution of dissolved chemical species in groundwater on Saturna Island can be explained by a period of submergence followed by rapid re-emergence at the end of the Pleistocene”.

The simulations involve a two step process:

- Determining the amount of intrusion after a period of submergence of 500 years (each scenario was run for a total of 1000 years to determine the effect of a longer period of submergence). The purpose of these simulations is to determine the amount of saltwater that intruded the island while it was partially submerged, and also to determine “initial concentrations” for the rebound simulations.

- Determining if the present day distribution of Cl (as measured by TDS) is consistent with that predicted for an evolution to present time given the initial condition prescribed by the above simulation.

If the proposed mechanism is correct, then two conditions should be met:

- In relation to process 1 above, saltwater must be able to intrude the aquifer to a sufficient degree and in a prescribed time period of 500-1000 years to result in NaCl water being present inland and at moderate to high elevation.

- In relation to process 2 above, the current distribution of Cl should be consistent with that predicted by the paleohydrogeology.
A second objective of this chapter is to use our understanding of the evolution of a long time process to evaluate the hydrogeologic parameters assigned to the model. Simulations described in Chapter 4 indicated that it takes approximately 1000 years to achieve a steady-state position of the freshwater-saltwater interface from a "cold" model start. Given that we know the approximate position of the interface today, it may be possible to use a representative Quaternary history to provide additional constraints on the hydraulic properties that describe the aquifer.

In the following sections, the Pleistocene history of Saturna Island is described and interpreted within the context of prescribing model boundary conditions that might serve to test the hypothesis.

**The Pleistocene History of Saturna Island**

The sea level history of the study area is important because sea level controls the amount of saltwater intrusion in coastal aquifers. The latest major glaciation that affected southwestern B.C. was the Late Wisconsinan, Fraser Glaciation ca. 30-10 ka (Clague, 1994). At the maximum of this glaciation the Cordilleran Ice Sheet covered British Columbia, Yukon, and Southern Alaska and extended to Puget Sound (Clague, 1994). The ice sheet developed in the high areas of the Coast Mountains and extended across the entire coast of British Columbia achieving a thickness of 2000 m. However, growth and decay was not uniform and, consequently, the sea level history is complex.

**Glaciation Style**

A scenario for ice sheet development was proposed by Davis and Mathews (1944) who identified four phases: an alpine phase at the beginning of the glaciation, an intense alpine phase when the ice lobes coalesced and covered the mountains, a mountain ice sheet phase during prolonged cold periods, and a continental ice phase when ice thickened and covered the entire region. At the beginning of the Fraser glaciation, approximately 27,000 yrs ago, the glaciers were restricted to high elevations (Figure 5.1). Ice flow was influenced by topographic relief; glaciers flowed in a
Figure 5.1. Growth (a) and decay (b) of the Cordilleran ice sheet in southern British Columbia during the Fraser Glaciation (from Clague, 1989, permission from Clague, 2003).
complex fashion to lower elevations, as evidenced by striations and ice flow indicators (Clague, 1989). Periods of growth were interrupted by recession or stabilization, which were controlled mainly by climate changes or eustatic sea level. Nourished by moisture from the Pacific Ocean, the ice sheet advanced from the high mountain areas, and extended to low areas, as shown in Figure 5.1, reaching the continental shelf of Vancouver Island, the Juan de Fuca Strait, and Puget Sound by 14,500 years BP (Porter and Swanson, 1998). Ice built up rapidly, especially during the climatic Vashon Stade (18,000-12,000 C\(^{14}\) years) of the Fraser Glaciation. Thus, in less than 4,000 years, mountain ice sheets had coalesced to form a continuous continental ice sheet that covered the entire province. At its maximum extent, the ice sheet had a maximum thickness of 2000 m.

Ice sheet decay was much more rapid than ice sheet growth. Retreat began at the continental shelf, and proceeded eastward and northward. Near the end of the Fraser Glaciation glaciers were active, and were restricted to valleys and fjords. In less than 1,000 years after the beginning of deglaciation the present day Vancouver and Victoria were ice free. Lowlands were free of ice 12,500-13,000 \(^{14}\)C years ago, and by about 9,500 years ago glaciers had the same extent as they do today.

**Sea Levels**

As an ice sheet advances, surface loading increases and causes isostatic depression and crustal deformation in a region. Land depression is typically more intense where the ice sheet is thicker. Ice retreat is accompanied by rebound of the land due to the ice melting. On the west coast, isostatic depression advanced towards the coast and lowlands, but did not have the same intensity in all areas due to variations in ice sheet morphology.

A consequence of ice loading is relative sea level variation. In general, sea level is lower when the land is not covered by ice and higher when the land is depressed. However, sea level changes are also influenced by other factors. The combined effect of isostasy, eustasy, and diastrophism must be considered. The effect of each factor on sea level evolution is summarized in the following section.
Factors that Control Sea Level

Isostasy is crustal adjustment to surface loading. Such adjustments have occurred globally during and following each glaciation as a consequence of ice loading, but they have been greater at the periphery of, and beneath, large continental ice sheets (e.g., British Columbia, Walcott, 1970). The growing mass of ice depressed the land through accommodation by plastic flow of the mantle. Total isostatic depression due to the Cordilleran ice sheet is estimated to be on the order of 300 m at Vancouver (Clague, 1994). When the ice sheet melted, the land rebounded causing isostatic uplift.

A model by James et al. (2000) for the southern BC coast assumes that the Late Wisconsinan Cordilleran ice sheet started its growth 25,000 years BP and attained its maximum extent 14,000 years BP. The model also assumes that the ice sheet decayed rapidly 12,000 years BP, and disappeared 10,000 years BP. Model results suggest rapid deglaciation, between 12,500 and 11,500 years ago, when a rapid rise of about 24 m in eustatic sea level occurred (Fairbanks, 1989). However, this model explains only about 20% of the total rise in relative sea level measured for the region.

Isostasy has two components; a hydrostatic one and a glacioisostatic one.

- Hydroisostasy is represented by the redistribution of weight on continental shelf due to sea level changes because of glaciation. Removal of water from the oceans reduce the load on ocean crust causing upward isostatic adjustment and sea level rise, whereas adding water to the oceans increases the load on the ocean crust causing downward isostatic adjustment and sea level fall.

- Glacio-isostasy is depression of the crust due to the ice sheet load. During each glaciation a large amount of ice was trapped on the land. The amount of depression varies with glaciation phase as well as with location. Previous isostatic studies have shown that upper mantle viscosities are lower in tectonically active zones than in the interior of tectonic plates (James and Clague, 2000). Therefore, a higher amount of glacioisostatic depression can be anticipated in coastal and lowland areas of BC. Sea levels throughout the region evolved from a high stand when the ice started its retreat, to a low stand some 9,500 years ago when sea
level was below today's value. At that time the eustatic component started to become the controlling component of sea level variation.

**Eustasy** - refers to worldwide changes in sea level as a consequence of the changing volume of glacier ice on land. During the Pleistocene there were numerous minor and major changes in ice volume, and global sea level adjusted itself accordingly. At the maximum phase of glaciation, sufficient water was locked up in the continental ice sheet to lower sea level by 125-130 m (Fairbanks, 1989).

**Diastrophism** - Sea level may be influenced by tectonism in this area. The Strait of Georgia is located at the intersection of four major tectonic plates - Pacific plate, Explorer plate, Juan de Fuca plate, and North American plate. The Strait of Georgia is a northwest – southeast orientated topographic forearc depression separating Vancouver Island from the mainland. Seismicity within the Strait of Georgia itself is low in comparison to more active areas to the south in Puget Sound. Nevertheless, geological evidence, consisting of buried marsh soils abruptly overlain by tidal muds, tsunami deposits and liquefaction features, support the fact that the region as a whole has experienced numerous large magnitude earthquakes in the past.

**Relative Sea Level Curves**

Sea level evolution for different areas on the west coast of BC was determined by Clague (1989) and James et al. (2000) based on studies on the Cordilleran Ice Sheet, terrestrial and marine sediments, and landforms. Relative sea level curves are presented here for specific parts of the BC coastal region and are used to estimate a sea level curve that might be appropriate for Saturna Island.

For ease of interpretation, the southwestern area of Canada was divided into three zones (Clague, 1989):

- **Inner coast**, which includes the mainland and adjacent islands,
- **Middle coast** (the area of concern), which includes eastern Vancouver Island and Gulf Islands,
- **Outer coast**, which consists of the Queen Charlotte Islands and western Vancouver Island.
Figure 5.2 shows the sea level evolution for the last 13,000 years, for different locations in southwest BC. Relative sea level curves were determined for two adjacent locations (Clague, 2002) to Saturna Island; the area of concern (middle coast) is approximated to be somewhere in between these two.

The middle coast, exhibited a high relative sea level during deglaciation and a Late Pleistocene marine limit of 75-175 m elevation (Clague, 1981). For the Gulf Islands region, sea level may have risen to cover the islands to an elevation of around 100 to 150 m.a.s.l. Thus, only the highest elevations on Saturna Island may have been above sea level (Figure 5.3). It is important to note that detailed studies of the southern Gulf Islands have not yet been undertaken, and the elevation of submergence is unconstrained.

The evidence that supports the sea level history scenarios includes radiocarbon ages for delta sediments at Courtenay and Parksville, located on the eastern part of Vancouver Island (Clague, 1980). At Courtenay, the remains of a high delta indicate that sea level fell from 150 m at 12,500 years BP to 21 m at 12,200 years BP. For Parksville, sea level was situated at 106 m above present elevation at 12,000 years BP, and fell rapidly to 52 m. At Portage Inlet, situated southwest of Saturna Island, the fibrous peat that accumulated 1-2 meters below present sea level between 5,000 and 9,250 years ago suggests a low sea level for this period. The peat is underlain and overlain by marine mud resulting from high sea level at 11,700 years BP and 5,470 years BP, respectively, when sea level rose.

All evidence suggests low sea level between 5,000-11,000 years BP. Before this period, sea level was higher at 75 to 175 m elevation, depending on location. As deglaciation started, about 13,000 years ago, sea level started to drop as an effect of the isostatic rebound.

In the two peripheral areas, sea level started to drop 13,000 years ago. The rate of drop was higher on the outer coast as the glaciers retreated from these areas. Isostatic rebound dominated during this time and sea level was almost 11 m lower than present elevation by 10,000 years BP (Clague and James, 2002).

To summarize, as a consequence of sea level rise, saltwater likely intruded the Gulf Islands, as the islands were, at least, partially submerged. Following
Figure 5.2 Shoreline evolution for different locations in Southwest B.C. over the last 13000 year (adapted after Clague, 2002).
Figure 5.3 Shoreline evolution during the last 12500 years on Saturna Island. The island is submerged 12500 years ago when sea level was approximately 150m above the actual position, followed by rapid rebound 11500 yrs ago.
rebound, seawater was displaced by freshwater; a consequence of the influx of fresh
rainwater.

**Transient Simulations for Submergence**

Based on the findings of steady state numerical simulations carried to investigate
the factors controlling the interface position, transient simulations are carried out to
model the evolution of saltwater intrusion over the last 12,000 years.

The conceptual model for Saturna Island was modified to incorporate sea level
changes as reflected in changes to the boundary conditions. All the other parameters, as
described previously, were kept the same. Recharge was assumed to be similar to
present values due to lack of historic information; however, sensitivity analysis showed
that the model is not particularly sensitive to the amount of recharge. Specified pressure
boundaries are used to represent sea level elevation 12,000 years ago. For the
purposes of the simulations, it is assumed that sea level was 150 m above its current
elevation. For two different scenarios, the steady state TDS concentrations are
compared in an effort to determine how long it takes for saltwater to intrude the island.

**Scenario 1**

In the first scenario all the parameters were kept the same as in Scenario 1 from
the previous chapter, and encompass low–range permeability layers, consistent with
measured values. Sea level, expressed as a specified pressure boundary was raised to
150 m. Sea level was kept constant at this elevation for 1000 years, because the main
purpose of this simulation was to determine how much saltwater would intrude the island
under submergence conditions.

A time series of TDS contour maps are provided (Figures 5.4 to 5.5). These
maps illustrate the possible evolution for 1000 years of submergence. After 500 years of
submergence (Figure 5.4), saltwater intrudes the island above its present level. The
0.0357 kg/kg TDS line advanced slowly to the centre of the island over this
Figure 5.4 TDS concentrations after 500 years of submergence of 150 meters below the present sea level

Figure 5.5 TDS concentrations after 1000 years of submergence of 150 meters below the present sea level
time period, to 300-400 m below ground surface. After 1000 years of submergence (Figure 5.5) the rate of saltwater intrusion increased in the northern part of the island, where higher permeability layers are present; however, the position of the interface (marked by the 0.005 TDS equiconcentration line) is situated 250-300 m below ground surface. Freshwater occurs at higher elevations beneath Mount Elford, as a consequence of the presence of highly fractured mudstone layers. Saltwater is observed to intrude the system faster at the ocean boundaries.

Based on the sea level history, relative sea level dropped after 500 years of submergence. Thus, the concentration distribution after 500 years of submergence was used as initial concentration conditions for a transient simulation to determine the effect of rebound on the movement of the interface.

The gradual decrease of sea level over this short rebound period could not be simulated because numerical instabilities occurred and the model did not converge to a solution. Therefore, a step decrease in sea level had to be used. This instantaneous drop in sea level is necessary approximation, but is consistent with reported “rapid” rebound in the region. Therefore, sea level was dropped from 150 m to its present level and the model was run until it reached steady-state. Note that sea level actually dropped to a sea level below present day elevation and gradually recovered to its present level over a period of 10,000 years. This gradual recovery was not included in the model because the difference in elevation is minimal.

Figure 5.6 depicts the TDS concentrations after 3000 years, while Figure 5.7 depicts the TDS concentrations after 6000 years. The emergence period was simulated in two steps of 3000 years to avoid numerical oscillations. The concentration distribution at the end of the first 3000 years was used as an initial condition for the second 3000 year simulation.

Steady-state is reached after 6000 years. Freshwater occurs 350 m below the ground surface at the centre of the island. A comparison between calibration curve for present day concentrations and “the Pleistocene model” results is shown in Figure 5.8. A 1-2 order of magnitude difference is evident between observed and simulated TDS concentrations; however, the results of “transient” and “steady-state” simulations are similar indicating that the submergence-rebound theory it is likely valid for this area.
Figure 5.6 TDS concentrations after 3000 years following the island rebound (corresponds to 9500 years ago).

Figure 5.7 TDS concentrations after 6000 years following the island rebound (correspond to 6500 years ago) – the system has reached steady-state.
Figure 5.8 TDS concentration comparison between the calibration curve and the rebound simulation at steady-state for Scenario 1.
Scenario 2

In this scenario, a higher range of permeability is used for the simulation (Scenario 2, Chapter 4). A period of submergence of 500 years was considered.

After 500 years (Figure 5.9) there is a higher magnitude of intrusion, compared with Scenario 1 as marked by the presence of saltwater to depths 100-200 m below ground surface. A longer simulation, for a period of 1000 years (Figure 5.10) results in a greater degree of intrusion, due to the higher permeability values used in this model. Freshwater is confined to the upper portions of the aquifer (150-200 m in the centre and southern areas). On the northern part of the island freshwater is present at greater depth due to the presence of higher permeability layer that dips towards north.

Two time steps of 3000 years each were used to arrive at a solution. After 3500 years (corresponds to 9500 years ago) freshwater flushes the saltwater rapidly saltwater from the upper layers, and the interface is "pushed" 100-150 m below actual sea level (Figure 5.11). After another 3000 years (Figure 5.12) the system reaches steady-state.

The steady-state TDS concentrations are compared with the TDS calibration curve in Figure 5.13. The same difference between observed and simulated TDS is evident as for the "steady-state" simulations. TDS concentrations are up to an order of magnitude below the calibration curve.

Nevertheless comparing the results of both scenarios with the "steady-state" TDS curves shows that the time frame is correct, the same position and shape of the interface is obtained using either steady-state or transient conditions.

In an effort to more closely match the TDS calibration curve to the simulated concentrations, simulations were attempted using higher values of permeability (one order of magnitude higher than scenario 2). However, unfortunately, the model would not converge and the effort was abandoned. It is expected that slightly higher permeability values may improve the calibration results.
Figure 5.9 TDS concentrations after 500 years of submergence of to a level of 150 m above present day sea level (Scenario 2).

Figure 5.10 TDS concentrations after 1000 years of submergence to a level of 150 m above present day sea level (Scenario 2).
TDS concentrations after 3500 years

Figure 5.11 TDS concentrations after 3000 years of simulations following island rebound considering a submergence period of 500 years (corresponds to 6500 years ago).

TDS concentrations after 6500 years

Figure 5.12 TDS concentrations after 6000 years following island rebound considering a submergence period of 500 years (corresponds to 9500 years ago) - the system has reached steady-state.
Figure 5.13 TDS concentrations comparison between calibration curve rebounded island scenario 2 (500 and 1000 years of submergence).
CHAPTER 6 – SUMMARY AND CONCLUSIONS

The two main objectives of this thesis were first, to determine the effect of fracturing on hydraulic testing and test the applicability of the equivalent medium concept by modeling of saltwater intrusion in a regional setting, and second, to determine if the Pleistocene sea level history for the region can account for the present day distribution of saline ground water beneath Saturna Island. Both objectives were achieved in the following steps:

- Analysing pump test data and interpreting the results within the context of the local geology;
- Running steady-state simulations to calibrate the model. Calibration involved varying the hydraulic parameters to approximate the present day position of saltwater-freshwater interface;
- Running transient simulations to determine whether the present day saltwater intrusion is a consequence of, or can be explained by, the Pleistocene history.

Pumping test data from the entire Gulf Islands region were used characterize both the geological medium and groundwater flow. As most of the tests where conducted near fracture or fault zones, the results may be inferred to be biased towards higher values. The effect of fracture proximity was minimized by using the derivative method to select only that portion of the test results that represent radial flow. However, hydraulic conductivities resulting from these analyses varied over a large range. The calculated hydraulic conductivity values did not show any consistent trend that would suggest higher values are associated with mudstone dominant formations and lower values with sandstone dominant formations, as previous geologic, hydrogeologic and geophysical studies would suggest. However, heterogeneity resulting from fracturing and the limited amount of data available are insufficient to make any strong conclusions in this respect.
A structure- and lithology-based conceptual model was used to simulate the present position of saltwater intrusion. The equivalent porous medium concept represented the framework for the numerical model. Three 2-D cross-sections were developed to simulate saltwater intrusion. A sensitivity analysis was undertaken to model the effects of: heterogeneities represented by layering, variation in recharge value, and applying different boundary conditions. Each formation present on Saturna Island was represented by permeabilities that are thought to represent the degree of fracturing present. An attempt was made to calibrate the model to observed concentration variations with depth; however, calibration was difficult due to poor spatial representation of the groundwater chemistry.

The first model across East Point peninsula showed that layering plays an important role in groundwater flow and solute transport. The best match between observed concentrations and model concentrations was obtained for the three layer scenario, which takes into account the presence of mudstone-dominant and sandstone-dominant formations as well as the presence of highly fractured mudstone-sandstone interlayers. However, the results indicated that permeability values should be increased relative to average values determined from Theis to reproduce the observed TDS concentration increase with depth.

A model representing East Point along the strike of the peninsula showed that layering controls the amount of saltwater that intrudes the island. This model reproduced the actual configuration of the saltwater intrusion only for the first 50 m below the ground surface indicating that the major pathways for intrusion are preferentially on the north and south sides of the island, perpendicular on the orientation of this model. Thus, this model is indicative of major flow directions along East Point peninsula.

Cross-section models across the entire island were developed considering the previous model findings. The geology of Saturna Island was represented in the model as an equivalent porous media, but the model did not account for the presence of mudstone-sandstone interlayers because incorporating these resulted in numerical instabilities. The results of the three simulations presented in this thesis indicate that a higher range of permeability values than the calculated average permeability values is necessary to simulate the observed concentration values. These higher values may be anticipated due to the fact that permeability values were calculated from transmissivity.
values divided by the total depth of the open borehole, rather than the sum of the fracture apertures along the borehole that contribute to flow.

Although higher permeabilities values appear to better reproduce the assumed “actual” distribution of the total dissolved solids concentration with depth, only the first 2 scenarios were used to simulate transient flow and transport across the island. The transient models were constrained by the Pleistocene history for southwestern Gulf Islands, in which the islands were partially submerged (to roughly 150 m above present day sea level) under seawater for 500 years. Submergence was followed by rapid emergence. Transient simulations showed that under submergence conditions it takes more than 1,000 years to “fully” saturate the island with saltwater.

Considering the similarity of the results of “steady-state” and “transient” simulations, the time frame and the amount of submergence are sufficient to allow the freshwater to replace the Pleistocene saltwater. Considering the degree of uncertainty in the permeability values, the amount of recharge and the actual sea level history the TDS concentrations for the transient model appear to agree well with the measured values.

While these models appear to represent, in general, the configuration of the interface at a regional scale, local variations in the interface, brought about by fracturing at the local scale, cannot be adequately represented using an equivalent porous media approach. Recognizing the important role of fractures as providing conduits for entry of saltwater into a freshwater aquifer, future work could aim to model solute transport flow through discrete fractures. This would require a large number of measurements, in terms of fracture orientation, spacing, aperture, etc.; data that are difficult to collect at any more than a local scale. Even if such data were available, SUTRA code would not be a suitable code to use for this type of modeling study. A discrete fracture, density-dependent flow and solute transport model would be required.

Improvements to the database include the following:

- Fracture measurements are needed to obtain a better representation of the hydraulic parameters.
- More groundwater chemistry data are required to describe the actual position of the interface.
Accurate water table measurements and/or better estimates of recharge are needed to better define the recharge boundary.

A more accurate time frame during the sea level history during the Pleistocene and to present day would be beneficial for constraining the model boundary conditions.


Haro Strait, Boundary Pass and Satellite Channel Map, 1996. Published by The Canadian Hydrographic Service, Scale 1:40,000, Minister of Fisheries and Oceans Canada.


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Appendices

Note: Appendices A, B, C and D are enclosed on a C.D. attached at the end of this thesis; a short description regarding the organization of each appendix is included as following.
Appendix A: Pumping test raw data, Log-log graph, Semi-log graphs, Derivative Method and Derivative Method Graphs

Appendix A contains raw data from the pumping tests conducted on Gulf Island region. The information is organized in two folders:

- Long-duration tests
- Short duration tests

Each test includes pumping and recovery data for pumping and observation wells.

1) Pumping test files are organized in 6 pages describing the data obtained from each test, the aquifer properties obtained by analyzing pumping tests data, log-log graphs of time versus drawdown, semi-log graphs of time versus drawdown, derivative method data and derivative method graphs.

2) Recovery test files are organized in three pages that include raw data, log-log graph and aquifer properties obtained by analyzing the data.
Appendix B: Aerial Photos

Appendix B contains aerial photos of Gulf Islands, showing the location of the pumping tests conducted in the region, and it is organized in 6 five files as following:

Figure B.1 – Galiano Island.
Figure B.2 – Gabriola Island.
Figure B.3 - Mayne Island.
Figure B.4 – Saturna Island.
Figure B.5 – Denman Island.
Figure B.1  Galiano Island.
Figure B.2. Gabriola Island
Figure B.3. Mayne Island.
Figure B.4 Saturna Island.
Figure B.5 Denman Island.
Appendix C: Long - Duration Tests description

Appendix C contains a description of each site for long duration pumping tests (where information was available).

Along with the description of geology of the area, in this appendix are included information on pumping rate changes, short description of the response of the aquifer to the test, log-log, semi-log and derivative graph and a summary of the results obtained for each method. This appendix also includes few comments on the site particularities and where aerial photos were available it integrates the results in the geological context.
Appendix D: Short - Duration Tests Description

Appendix D contains aerial photos showing the location of each test along with log-log, semi-log and derivative graphs.