MODELLING THE IMPACTS OF CLIMATE CHANGE ON GROUNDWATER: A COMPARATIVE STUDY OF TWO UNCONFINED AQUIFERS IN SOUTHERN BRITISH COLUMBIA AND NORTHERN WASHINGTON STATE

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ABSTRACT

A methodology is developed for linking climate and groundwater models to investigate future impacts of climate change on groundwater resources using two case study sites of unconfined aquifers in southern British Columbia and northern Washington State. One semi-arid site is compared with one wet coastal site. The two groundwater systems differ in river-aquifer interactions, recharge, aquifer heterogeneity, scale, and groundwater use. Climate change scenarios from the Canadian Global Coupled Model 1 model runs for 1961-2000, 2010-2039, 2040-2069 and 2070-2099 are downscaled to local conditions, modelled at daily time scales using a stochastic weather generator, and applied to the spatially-distributed infiltration model. At one site the basin-scale runoff is also downscaled to predict river discharge and river-aquifer interactions in future climates. The impacts of predicted climate change on the groundwater system for each site are modelled in three-dimensions using Visual MODFLOW. Results and methodologies are compared and discussed.

EXECUTIVE SUMMARY

A methodology is developed for linking climate and groundwater models to investigate future impacts of climate change on groundwater resources. Two case study sites are used to develop and test the methodology, as well as to compare the results between two climate regions. The Grand Forks aquifer in the semi-arid south-central interior of British Columbia (BC), Canada, is compared with the Abbotsford-Sumas aquifer in the wet coastal region of southwestern BC and northwestern Washington State, USA. The two groundwater systems differ in river-aquifer interactions, recharge, aquifer heterogeneity, scale, and groundwater use, but are both unconfined, heterogeneous, and highly permeable. In Grand Forks, the river-aquifer interactions dominate the hydraulic response. Climate change scenarios from the Canadian Global Coupled Model 1 (CGCM1) model runs for 1961-2000, 2010-2039, 2040-2069 and 2070-2099 are downscaled to local conditions using the Statistical DownScaling Model (SDSM). The predicted mean temperature increase, the changes in monthly mean precipitation, and the associated changes in wet and dry spells, are realistically modelled at daily time scales using the LARS-WG stochastic weather generator. Spatially distributed recharge is considered using surface and subsurface data, and the generated weather used to drive the recharge model. CGCM1 downscaling is used to predict basin-scale runoff for the Kettle River upstream of Grand Forks. These results were converted to river discharge along river reaches. In future climate scenarios the hydrograph peak is shifted to an earlier

date, although the peak flow remains the same. The impacts of predicted climate change on the two groundwater systems are modelled in three-dimensions using Visual MODFLOW, and the results compared. Direct impacts were represented as mapped changes in groundwater levels and as effects on groundwater-surface water interactions. Results suggest small, changes in groundwater levels, forced by changes in recharge. At Grand Forks, water levels within the floodplain respond significantly to shifts in the river hydrograph under scenarios of climate change than to climate-caused changes in aquifer recharge. The Abbotsford-Sumas aquifer is mainly recharge controlled and subsequent groundwater level responses to climate change predictions are relatively small, but are highly variable in space and may affect baseflow discharge in streams draining the aquifer.

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GLOSSARY

BRANCH	One dimensional river flow model (USGS model)
CGCM1	Canadian Global Coupled Model 1
CICS	Canadian Institute for Climate Studies
GCM:	Global Climate Model or General Circulation Model
GIS	Geographic Information Systems
GMS	Groundwater Modelling System (modelling software)
HELP	Hydrologic Evaluation of Landfill Performance (US Environmental Protection Agency)
IPCC	Intergovernmental Panel for Climate Change
LARS-WG	One type of stochastic weather generator
MODFLOW	Modular Groundwater Flow Model (USGS model)
NCEP	National Centre for Environmental Prediction
PCA K-nn	Principal Component Analysis k-means method
SDSM	Statistical Downscaling Model
WG	Weather Generator
WGEN	One type of weather generator

1 INTRODUCTION

Background

In the studies of the potential impacts of global climate change on ground water resources, the interactions between unconfined aquifers, rivers, and the atmosphere are modelled to determine impacts on water levels, and also how groundwater resources are affected by climate variability and climate change, to improve forecasts of future impacts. It is expected that changes in temperature and precipitation will alter recharge to groundwater aquifers, causing shifts in water table levels in unconfined aquifers as a first response to climate trends [Changnon et al, 1988; Zektser and Loaiciga, 1993]. Although the most visible impacts could be changes in surface water levels in lakes [Winter, 1983], a concern of water management and government officials is the potential decreases of groundwater supplies for municipal and agricultural use. Such changes might decrease quantity, and perhaps, quality of water, which would also have detrimental environmental effects on fisheries and other wildlife by changing baseflow dynamics in streams [Bredehoeft et al, 1982; Gleick, 1986].

Aquifer recharge and groundwater levels interact, and depend on climate and groundwater use. Each aquifer has different properties and requires detailed characterization and, eventually, quantification (e.g., numerical modelling) of these processes, and linking the recharge model to climate model predictions [York et al, 2002]. Large regional and coarse-resolution models have been undertaken to determine the sensitivity of groundwater systems to changes in critical input parameters, such as precipitation and runoff [York et al., 2002, Yusoff et al., 2002], with few exceptions of very small aquifers and detailed investigations [e.g., Malcolm and Soulsby, 2000]. Of particular interest are coupled hydrologic systems, where changes in surface flow regime and changes in groundwater recharge interact to affect both groundwater and surface water levels.

The overall uncertainty of climate change scenario simulations in aquifers depends largely on the spatial and temporal resolution of such models. In practice, any aquifer which has an existing and calibrated conceptual model, together with calibrated numerical model, can be assessed for climate change impacts through scenario simulations. The accuracy of predictions depends largely on the scale of the project and the availability of hydrogeologic and climatic datasets and the level of uncertainty of both climate model simulations and downscaling of such simulation results to local conditions.

Significance of this Study

This study was motivated by the Canadian government's efforts for assessment of impacts of climate change and climate variability [Environment Canada, 1997]. In 1996, Environment Canada initiated a countrywide study to evaluate these impacts on Canada as a whole, and to consider existing and potential adaptive responses [Environment Canada, 1997]. Impacts on hydrologic systems are expected to be significant in most parts of Canada and, specifically, in British Columbia (BC) where groundwater management is among the important water issues facing the province [Environment Canada, 2000]. While not pervasive in all regions of the province, evidence of limited water availability exists, especially in the southern interior of the province. For instance, over 17% of surface water sources are at, or nearing, their capacity to reliably supply water for extractive uses [BC Ministry of Environment, Lands and Parks, 1999]. Groundwater-surface water conflicts have been identified in a few interior aquifers. Knowledge of groundwater flow paths is also required for bi-national management of contaminants and water supplies, especially in the Abbotsford-Sumas aquifer in the lower Fraser Valley.

Water resources are central to any study on climate change; however, most research to-date in BC has been directed at forecasting the potential impacts to surface water hydrology [e.g., Whitfield and Taylor, 1998]. Relatively little research has been undertaken to determine the sensitivity of groundwater systems to changes in critical input parameters, such as precipitation and runoff, despite the fact that BC is one of the largest users of groundwater in Canada. In the south-central interior of the province, where agriculture is a significant component of the economy, groundwater resources may be particularly impacted directly and indirectly by climate change. The purpose of the current research study, then, is to model the sensitivity to climate change, and to identify any potential impacts of climate change on the unconfined aquifers in southern BC.

Through this research, we applied several existing methods for linking climate change predictions for regional aquifers, and developed new procedures for analysis and interpretation of modelling results and dealing with uncertainties. To accomplish a realistic link between regional climate, and station-specific climate, and the groundwater system at an appropriate scale, two regional aquifers (less than 150 km^2 in area) were selected to test the high resolution groundwater flow models, climatic inputs through recharge, and climate-driven surface water links, where appropriate. The two aquifers were the Grand Forks aquifer in south-central BC, and the Abbotsford-Sumas aquifer in south-western BC and northern Washington State, USA. In essence, this is a comparative study of two aquifers and effects of predicted site-specific climate change on groundwater resources.

The approach consisted of constructing a three-dimensional groundwater flow model for each aquifer, created with the appropriate conceptual representation of the aquifer architecture (hydrostratigraphy) – see chapter 2 for more details. Careful consideration was given to direct recharge of the unconfined aquifers from precipitation, which involved the modelling of spatially distributed and temporally varying recharge, for each climate scenario separately. The latest developments in spatially-distributed recharge modelling [Jyrkama et al., 2002] and hydrostratigraphic modelling [Herzog et al., 2003] methodologies were incorporated. The goal was to permit a more comprehensive evaluation of water budgets, incorporation of seasonal changes in demand for groundwater, leading to a better understanding of the direct impact of climate change on alluvial aquifers.

The numerical models were developed based on critical review of all available information. Modelling methodology for groundwater was based on that described in published literature [e.g. Anderson and Woessner, 1992]. Model boundary conditions, especially rivers, were considered, where appropriate, and in as much detail as possible with the available information. The models were calibrated to historic water levels, then the various climate scenarios were input to the calibrated models. For the climate scenarios, recharge values for future climate change scenarios were modelled separately, and then input into the groundwater flow model, and the impact on water levels in the aquifer were calculated. The same methodology was used for both aquifer studies.

The aquifers studied in this research project will complement other intensely studied aquifers from other climatic regions, such as the Waterloo Moraine in Ontario [Martin and Frind, 1998]. This list will serve as the best candidates for modelling the impacts that might occur under future climate change scenarios. Furthermore, the same aquifers will be most likely studied in the future for other purposes, due to large datasets for hydrostratigraphy, hydraulic and water quality information, and the availability of calibrated numerical models.

Thesis Outline

This thesis is comprised of four papers. Three have been submitted to or are in press in peer-refereed scientific journals [Waters Resources Research (1), Journal of Hydrology (1), Global and Planetary Change (1)] and one is an extended version of a paper that was published in conference proceeding (Puget Sound-Georgia Basin Research Conference, 2005). The appendices consist of supplementary detailed reports submitted to federal and provincial governmental agencies (Environment Canada, Natural Resources Canada – Climate Change Action Fund, BC Ministry of Water Land and Air Protection), which funded this project.

The papers form the thesis chapters 2 through 5, with the exception of the abstracts and references. References were pooled together for the four papers, and form the thesis reference chapter. There are also final conclusions of the thesis, which draw on the conclusions of all the journal papers. Figure numbering is separate within each paper, although chapter number is included in the overall thesis figure and table numbering. Some text and figures, such as site maps, may be similar in some papers, but it should be noted that the papers were designed to stand alone when published. The following sections provide an overview of the content of these papers. The technical details of the modelling projects, including all methods and results, are included in Appendices 1 to 4 in electronic form on a CD (in pdf format).

Paper 1: Distributed Recharge Modelling

The first manuscript is entitled "Modeled Impacts of Predicted Climate Change on Recharge and Groundwater Levels", Scibek and Allen [submitted a]. This manuscript has been submitted to *Water Resources Research Journal.*

This paper describes the complete methodology and results of linking climate models and groundwater models to investigate future impacts of climate change on groundwater resources in the Grand Forks aquifer. Climate change predictions were obtained from Canadian Global Coupled Model (CGCM1) model runs for 1961-1999, 2040-2060, and 2080-2100. Data include absolute and relative changes in precipitation, including indirect measures of precipitation intensity, dry and wet spell lengths, temperature, and solar radiation. Climate data were downscaled to local conditions using two separate downscaling methods, SDSM and PCA K-nn, which were then compared.

The main uncertainty still lies in the downscaling method performance, as demonstrated with large calibration bias between the downscaled present climate and observed present climate at a particular location. In effect, we do not know the actual future climate at any of the study location, but we are estimating it with imperfectly calibrated downscaling models, from also uncertain results of CGCM1 climate model. At Grand Forks site neither SDSM nor K-nn adequately models precipitation, and the two models differed in goodness of fit to observed precipitation in different months of a year.

One of the problems is that precipitation is not represented directly in a GCM, only humidity and other meteorological variables of the atmosphere being modelled.

In contrast to precipitation, the air temperature variable is directly represented in a GCM, including CGCM1 output. The predicted changes in monthly mean precipitation, and the associated changes in wet and dry spells, were realistically modelled at daily time scales using the LARS-WG stochastic weather generator, and then applied as inputs to the HELP infiltration (recharge to aquifer) model. The resulting downscaling model bias to observed mean daily air temperature was much less than for precipitation, in both SDSM and K-nn downscaling models. The predicted mean daily temperature had an increasing trend in all months from present to future scenarios by $1^{\circ} - 2^{\circ}C$ per 30 years, which reflects the CGCM1 results for that geographic region (CGCM1 grid square).

Spatially-distributed recharge methodology is presented which accounts for unsaturated zone depth, aquifer heterogeneity, soil permeability, and irrigation return flow. Recharge was modelled using US EPA HELP (Hydrologic Evaluation of Landfill Performance) from 100 year weather runs for each climate scenario, averaged monthly, and input to each recharge zone separately. LARS-WG performance was evaluated, and the sensitivity of the recharge model to HELP parameters determined.

The spatial distribution of recharge was found to have consistently greater control on groundwater levels and flow rates than the temporal distribution of recharge in the groundwater flow model of the Grand Forks aquifer (dry climate), but it may be more significant in a wetter climate. If a mean annual recharge value is applied to the model, then modelled water levels are within 0.10 m of those calculated with temporally variable recharge applied at monthly intervals. The HELP model proved sensitive to several properties of the vadose zone; therefore, in order to achieve accurate results for recharge, the spatial variability of these key variables was considered in the development of recharge zonation maps for each study site. The improved resolution of recharge ensured that spatial distributions were accounted for in the analysis of climate change impacts, although the spatial distribution was of minor importance in the end at this particular aquifer.

At the Grand Forks site, overall the downscaled climate scenarios and the recharge model predicts that for the 2040-2069 scenario there will be 50% more recharge to the unconfined aquifer during the spring and summer seasons, compared to present climate scenario 1970-1999. In the autumn season, recharge is predicted to increase (10 to 25% depending on month within the season) or remain the same as present depending on location within the valley. In the winter, the CGCM1 predictions suggest less precipitation, and consequently, less recharge to aquifer. The predicted increase in recharge to aquifer will result, on average, in 0.2 m increase in groundwater elevation, although effects on future water supplies will be minimal.

The groundwater flow model also integrates transient river water levels as described in Scibek et al. [submitted], paper 2 / chapter 3 of this thesis. A transient three-dimensional groundwater flow model, implemented in Visual MODFLOW (ver. 3.1.84), was used to calculate resulting water table elevations. Head differences (between historical and each predicted future scenario) were computed at each time step and mapped in GIS. The effect of applying spatially-distributed recharge on water levels was also investigated.

The detailed methodology of recharge modelling as linked to climate models, was also described in a report by Scibek and Allen [2003 b], attached in Appendix 1, and the detailed description of the numerical model for the Grand Forks aquifer, and the results of climate change impacts modelling is provided in a report by Scibek et al [2004], and also a summary report by Allen et al. [2004] presented to Natural Resources Canada that documents both methodologies and results of the climate change impacts modelling at Grand Forks aquifer, both reports attached in Appendix 2.

Paper 2: Groundwater-Surface Water Interaction

The title of the second paper, based on the Grand Forks aquifer study, is "Groundwater-Surface Water Interaction Under Scenarios of Climate Change Using a High-Resolution Transient Model" [Scibek et al., submitted]. The manuscript has been submitted to the Journal of Hydrology.

This paper describes the methodology and results used to quantify the surface water hydrology of the Kettle River at Grand Forks and the response of the aquifer under scenarios of climate change to changes in surface hydrology. CGCM1 downscaling was used to predict basin-scale runoff for the Kettle River upstream of Grand Forks. Specifically, the river discharge hydrographs (predicted and base case) were converted to river discharge along Kettle and Granby River reaches using the one-dimensional river model BRANCH, which takes into account channel geometry. A river stage schedule was computed and input as specified head boundary conditions in the groundwater flow model for each climate scenario. Head differences within the aquifer were computed at each time step for historical and future climate scenarios, mapped in GIS, and discussed with references to river-aquifer interactions, groundwater storage, and water budgets.

With changing climate, the changes in surface water streamflows may locally affect groundwater flows in surficial aquifers. The Grand Forks aquifer is one example of such situation. Groundwater levels in the Grand Forks aquifer respond more directly to changes in the timing of basin-scale snowmelt events in the Kettle River, and the subsequent shift in the timing of the hydrograph, under scenarios of climate change, rather than to changes in recharge. Between 11 and 20% of the river flows from the river into the aquifer during spring freshet, and storage duration in the alluvial aquifer lasts 30 to 60 days. Hydrologic modelling under scenarios of climate change suggests that Kettle River peak flow is expected to occur at an earlier date in the year and the baseflow period is expected to be of longer duration and lower than at present. The hydrograph shift for the 2040-2069 climate scenario is larger than in the 2010-2039 climate scenario, resulting in an apparent decrease in groundwater levels by up to 0.5 m during the spring season. In areas furthest away from the river influence, the direct precipitation recharge begins to dominate the response to climate change.

Groundwater modelling and monitoring should be continued, and further scenarios, which make predictions on changes in water consumption and climate together, evaluated. The groundwater resources in the valley will not be affected significantly by these changes as long as the Kettle River maintains its discharge and supplies large quantities of recharge to the aquifer. Except near pumping wells, the aquifer groundwater levels cannot drop below the Kettle River water levels in the valley, even if there is limited direct recharge from precipitation to the aquifer. In the end, the future groundwater use in the valley is limited by the withdrawal of an acceptable percentage of Kettle River discharge, especially at its minimum discharge rate in the late summer.

The resolution and formulation of the model accommodates changes in the river discharge/stage under the various climate change scenarios. These changes, in turn, are captured through the model boundary conditions, which consist of nodes of specified heads. Thus, in areas with strong river-aquifer interactions, it is very important to adequately represent the aquifer heterogeneity in the groundwater flow model to accurately predict changes in groundwater levels. The method of conceptual representation of aquifer heterogeneity in the model (influencing connectivity to the river) has uncertainty. This uncertainty is propagated through the modelling process and has as much influence on resulting modelled water levels (0.5 m changes in head) as do the predicted impacts of climate change on water levels. Therefore, in order to reduce these uncertainties, the spatial resolution of the model must be increased, thereby necessitating better resolution of aquifer heterogeneity.

The detailed methodology of river modelling as linked to climate models is described in a report by Scibek and Allen [2003 a], attached in Appendix 3.

Paper 3: Climate Change Impacts on Groundwater in the Abbotsford-Sumas Aquifer

The third paper describes the numerical model and results for the Abbotsford-Sumas aquifer and is entitled: "<u>Modelled Climate Change Impacts in the Abbotsford-Sumas Aquifer, Central Fraser Lowland of BC, Canada and Washington State, US.</u>" This paper is an extended version of one that was published in the Proceedings of the Puget

Sound–Georgia Basin Research Conference, 2005 [Scibek and Allen, 2005 b] provided in Appendix 4.

This paper presents the results of climate change impacts modelling in the Abbotsford-Sumas aquifer. A three-dimensional transient groundwater flow model, implemented in Visual MODFLOW, was used to simulate three climate scenarios in one-year runs (1961-1999 present, 2010-2039, and 2040-2069) in order to compare recharge and groundwater levels to present. The same methodology was used as in the Grand Forks aquifer study [Scibek and Allen, submitted a, b]. The purpose of the study was to 1) test the methodology developed for Grand Forks at another site, and 2) to provide data that could ultimately be used to compare the responses of two aquifers to climate change.

At the Abbotsford-Sumas site, direct recharge is predicted to decrease by 5.6 to 6.3% relative to historic values under climate change for the 2010-2039 scenario. Greater decreases in recharge were predicted for the 2040-2069 climate scenario. The groundwater flow model results showed spatially-variable reduction in water levels ranging from 0.05 m to more than 0.25 m in most upland areas. In the 2040-2069 scenario, groundwater level declines were also on the order of 0.25 m in most upland areas. These lower water levels will result in a reduction in hydraulic gradients from recharge to discharge areas, and a consequent scaled reduction in groundwater discharge. The lowering of the water table in the uplands area will most likely decrease baseflow in the streams, which are fed mostly by seepage of groundwater. Lowland areas cannot be assessed because the model was constrained by specified head boundary conditions associated with major streams.

Aquifer heterogeneity was also important in the characterization and modelling of the Abbotsford-Sumas aquifer. Due to the significant heterogeneity of the aquifer, the nature of interaction between the aquifer and the numerous streams could not be determined at a local scale. The model could be re-calibrated if the representation of heterogeneities was improved. The locations of perched water tables should be investigated and the calibration data set modified. However, there are cost limitations and diminishing returns from collecting more data on hydraulic conductivity in many areas of the regional aquifer. Specific areas of interest should be identified and new data collected. In particular, the uplands near Abbotsford where groundwater flow model calibration was poor and where perched water tables are suspected to be present, and also along important streams draining south from the uplands west of Abbotsford. An improved understanding of groundwater chemistry and perhaps use of tracers to delineate capture zones would help to validate flow model results.

Further detailed investigations are required to measure the interaction of surface water and groundwater, through streamflow measurements and water level surveys, direct measurements of infiltration rates for different soil types to validate our recharge estimates, and better coupling of surface and groundwater in the flow models. Improvements to the model should consider changes in hydrology as a consequence to climate change, but more site-specific information on the streams and refinement of the model in those areas is needed. More detailed flow models will require much better surveying of stream channels and surveying of static water elevations in wells for the purpose of model calibration. As well, soil permeability and infiltration rate data should be collected from many points in the valley to verify the recharge rates modelled in HELP.

The detailed description of the conceptual and numerical model of the Abbotsford-Sumas aquifer and the modelling results is provided in a report by Scibek and Allen [2005 a], attached in Appendix 4. An extended paper [Scibek and Allen, 2005 b], based on the content of Chapter 3, documents the climate downscaling and recharge modelling for that aquifer, and a second report [Scibek and Allen, in prep] describes the results of the climate change impacts modelling on that aquifer.

Paper 4: A Comparative Study of Climate Change Impacts in Two Surficial Aquifers

The fourth and final paper is entitled "Comparing the Responses of Two High Permeability, Unconfined Aquifers to Predicted Climate Change", Scibek and Allen [in press]. The manuscript is to be published by the journal Global and Planetary Change.

The paper compares the results from climate change impacts modelling at the Grand Forks and Abbotsford-Sumas aquifers. These two case studies shared common methodologies throughout and this facilitated a comparison of the model results at all steps of the modelling process. The common methodology for downscaling climate model results to local conditions, and then using weather generation to drive the recharge model created a defensible and standardized methodology for generating recharge predictions for groundwater modelling projects.

At Grand Forks aquifer, the downscaling of GCM results were problematic, resulting in large bias for summer monthly precipitation as described in paper 1 in this thesis. However, at the coastal location of the Abbotsford-Sumas aquifer, the two downscaling models were much more consistent and had much smaller calibration bias to observed precipitation, and consequently, much greater confidence in precipitation predictions in future climate scenarios. At the interior location, summer precipitation was predicted to increase slightly, whereas at the coastal location summer precipitation was predicted to decrease slightly. The changes were not uniform at monthly time scales, but had seasonal differences.

As a result of all the uncertainties involved with climate downscaling, aquifer property distributions, and model calibration, at this time the results of this study should be treated as a sensitivity study rather than actual predictions, even though we attempted to use "actual" best scientific guesses (model results) at future climate predictions, linked to the groundwater model through documented and defensible methodology presented in this thesis. This work showed one method of linking the groundwater flow models to climate model outputs, and demonstrated that such links are practical to use in groundwater modelling studies, but that there are many uncertainties involved at each step. Climate model downscaling to one location has large uncertainty and should be improved in future studies. The spatial distribution of recharge may be important in aquifers where recharge from precipitation dominates all other recharge pathways. Aquifer heterogeneity representation is always important, and there is very large uncertainty involved when dealing with transient model behaviour, especially when river-aquifer interactions are important.

2 DISTRIBUTED RECHARGE MODELLING

Modeled Impacts of Predicted Climate Change on Recharge and Groundwater Levels

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Submitted to Water Resources Research [Scibek and Allen, submitted a]

Introduction

In studies of global climate change, the impacts on water resources and the interactions between unconfined aquifers and the atmosphere are studied and modeled to determine impacts on water table levels. It is expected that predicted global changes in temperature and precipitation will alter groundwater recharge to aquifers, causing shifts in water table levels in unconfined aquifers as a first response to climate trends [Changnon et al., 1988; Zektser and Loaiciga, 1993]. Most research to-date has been directed at forecasting the potential impacts to surface water hydrology, while for groundwater hydrology, typically only large, regional and coarse-resolution models have been undertaken to determine the sensitivity of groundwater systems to changes in critical input parameters, such as precipitation and runoff [e.g., York et al., 2002, Yusoff et al., 2002], with few exceptions of very small aquifers and detailed investigations of potential impacts of climate change (scenarios) on unconfined aquifer water levels [e.g., Malcolm and Soulsby, 2000].

Aquifer recharge has traditionally been difficult to estimate for large areas, but a variety of methods have been used [Simmers, 1998]; from statistical empirical models linking precipitation trends to aquifer recharge and groundwater levels [Chen et al., 2002], to spatially-distributed recharge applied to three-dimensional groundwater flow models [Jyrkama et al., 2002]. The validity of assumptions of recharge rates becomes very important in small-scale transient models, where detailed groundwater flowpaths and levels are required [Jyrkama et al., 2002]. For the purposes of climate change impacts modeling, relative changes in recharge rates are of interest, and particularly how these relative changes are translated to the groundwater levels. Ideally, recharge rates should be as accurate as possible to represent the small shift from present to future climatic conditions, particularly where recharge has dominant effect on the local water balance. In practice, however, data limitations preclude detailed and highly accurate estimates that can be applied over a large area. Thus, we are left to examine the sensitivity of recharge and groundwater levels to climate change using a systematic approach.

This study was motivated by the Canadian government's initiative to assess the impacts of climate change and to develop adaptive strategies for climate change under the auspices of Natural Resources Canada's Climate Change Action Fund. The goal of the study was to permit a more comprehensive evaluation of water budgets, and to provide a better understanding of the direct impact of climate change on unconfined alluvial aquifers. Other scientific research objectives were to evaluate the importance of spatial distribution of recharge on groundwater modeling results, and to identify uncertainties in the climate-to-model process. To answer these questions we present a methodology for linking GCM predictions (via downscaling) to a recharge model, and then to a groundwater flow model.

The selected case study area was the Grand Forks aquifer (34 km^2 in area), contained within the mountainous valley of the Kettle River near the City of Grand Forks, in south-central British Columbia (BC) along Washington State boundary (Fig. 1.). The climate is semi-arid and most rainfall occurs in summer months during convective activity. In the winter, much of the precipitation at high elevation is as snow, although the observing sites at valley bottoms record less snowfall. Groundwater is used extensively for irrigation and domestic use [Wei et al., 1994]. Within the Grand Forks valley, the Kettle River is a meandering gravel-bed river incised into glacial outwash sediments, and previous modeling studies [Allen et al., 2003] demonstrated that the aquifer water levels are highly sensitive to water levels in the Kettle River. Because climate change is anticipated to impact both the timing and amplitude of flow in the Kettle River, consideration of impacts of climate change must necessarily consider both surface water and groundwater in this aquifer [Scibek et al., submitted]; however, in this paper we focus on the effects of climate change on recharge from precipitation, touching on certain aspects of the hydrology where appropriate.

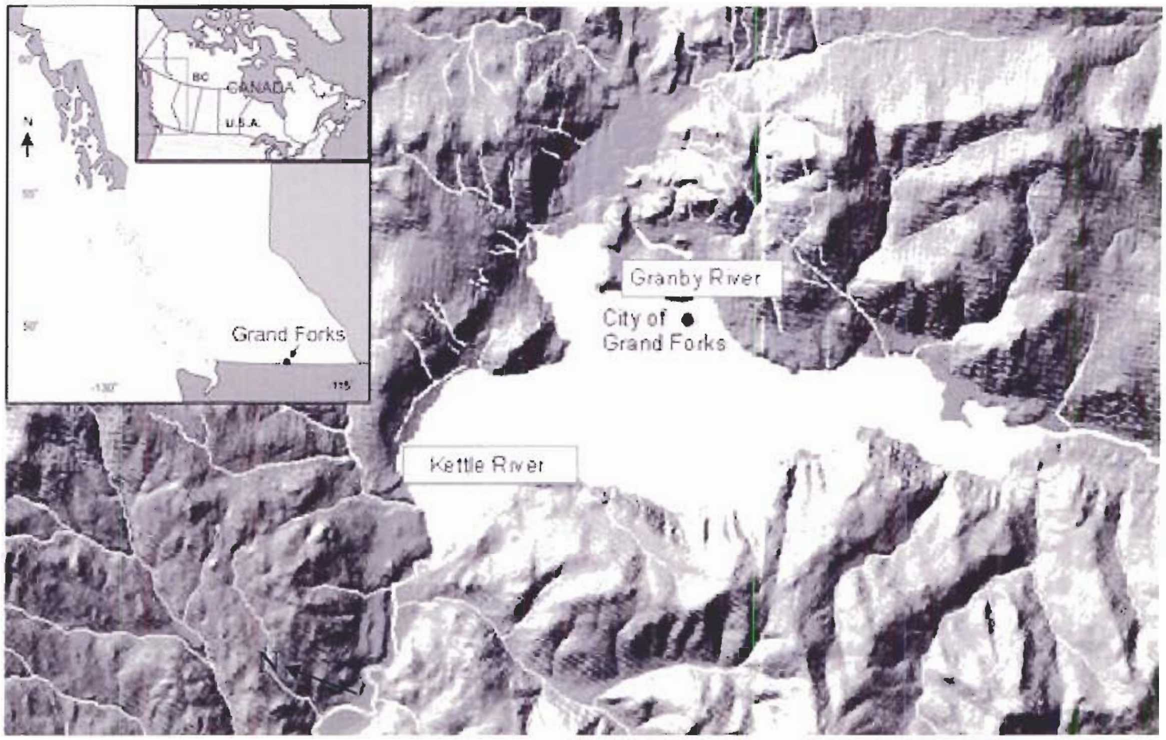

Figure (2) - 1: Mountainous topography of the Grand Forks valley showing the unconfined valley aquifer (filled grey) and drainage (white). The Kettle River (east-flowing) is shown to meander through the valley, and eventually discharges into the Columbia River. The valley widens near town of Grand Forks, where the Granby River flows into the Kettle River. Inset map shows the location map of the study area in British Columbia, Canada.

Downscaling of GCM Predictions

Spatial downscaling techniques [Hewitson and Crane, 1996; Wilby and Wigley, 1997] are used to derive finer resolution climate information from coarser resolution GCM output, assuming that the statistical relationships, linking observed time series to GCM variables, will remain valid under future climate conditions. GCMs do not accurately predict local climate, but the internal consistency of these physically-based climate models provides most likely estimates of ratios and differences (scaling factors) from historical (base case) to predicted scenarios [Loaiciga et al., 1996] for climatic variables, such as precipitation and temperature.

Climate scenarios for modeled present and future conditions were taken from the Canadian Global Coupled Model (CGCM1) [Flato et al., 2000] for the IPCC IS92a greenhouse gas plus aerosol (GHG+A1) transient simulation. CGCM1 predictions are valid for Canada and fall in the average of other GCMs. These include absolute and relative changes in precipitation and temperature. Temperature statistics were: mean, median, minimum, maximum, variance, and inter-quartile range. Precipitation variables were: mean, median, maximum, variance, dry/wet spell length, and % wet days in the month. Five daily data sets for CGCM1 were obtained from the Canadian Institute for Climate Studies [CICS, 2004] for a grid location nearest to Grand Forks (Y=11 Latitude: 50.09°N and X=16 Longitude: 120°W – Grand Forks is at 49.1°N and 118.2°W). Four were CGCM1 scenarios, each with data for a number of potential predictor variables. The "current climate" scenario was generated by CGCM1 for the period 1961-2000. The subsequent "future climate" experiments using CGCM1 with GHG+A1 were for 2020s, 2050s, and 2070s. The fifth data set was a calibration data set for the downscaling model. The calibration dataset contains observed daily data for 1961-2000, derived from the NCEP (National Centre for Environmental Prediction) re-analysis data set [Kalnay et al., 1996] for the period 1961-2000. Monthly means and other statistics were calculated from mean daily values, and the NCEP dataset had 10% or smaller bias to observed precipitation at Grand Forks (compared monthly means), thus we have high confidence in using NCEP data for calibration of downscaling model.

The downscaling of CGCM1 results was accomplished using two independently calculated methods: 1) Statistical Downscaling Model (SDSM) software [Wilby et al., 2002], and 2) principal component K-nn method [e.g., Zorita and von Storch, 1999; Yates et al., 2003] computed by Environment Canada [Whitfield and Cannon, 2000]. A more in-depth description of these methods and details of the comparisons are provided in Allen et al. [2004]; only a summary of the results is provide here. Four climate scenarios (30 years of daily weather) were generated using each calibrated downscaling model: current climate (1960-1999), 2020's climate (2010-2039), 2050's climate (2040-2069), and 2080's climate (2070-2099).

A comparison of the results from the SDSM and K-nn downscaling methods shows different magnitudes and directions of change, mostly for precipitation, in future climate scenarios, demonstrating the uncertainty associated with the process of downscaling. Although this uncertainty limits the predictive aspect of this (and similar) studies, it does not detract from the study's usefulness as a realistic sensitivity analysis to potential climate change, whatever the actual climate changes in each month will be in the future.

Downscaled daily temperature time series were analyzed for 1) mean, and 2) standard deviation. Temperature calibration results show that the two downscaling methods yield comparable estimates of mean monthly temperature, and calibration bias is small (Fig. 2). Similarly, both methods agree in the magnitudes and directions of temperature change, and represent an increase of approximately 1°C per 30 years for all months (Fig. 4).

Downscaled daily precipitation time series were analyzed for: 1) mean monthly precipitation, 2) standard deviation in daily precipitation, 3) % wet days, 4) dry series length, and 5) wet series length. Calibration results for both downscaling methods show variable results (Fig. 3). SDSM is better calibrated than k-nn in the wettest months of the year, although precipitation is underestimated by up to 40% in the summer months compared to observed (1961-2000 period). Precipitation has variable seasonal / monthly predicted changes, and results vary somewhat between downscaling methods. The change factors for precipitation, extracted from SDSM downscaled CGCM1 predictions, indicate an increase in July and August, variable changes (increase or decrease) in other months, and corresponding changes in % wet days for those months (Fig. 5). The relative changes in precipitation according to the downscaling results from CGCM1 model were graphed in Figure 5, separately for relatively dry and wet months of the year. The climate at Grand Forks is predicted to become slightly wetter in the "dry months" Feb-Mar and Jul-Aug (Fig. 5a), increasing to greater and greater amounts from present time to 2099. From present to the 2010-2039 period there is predicted increase in precipitation by factor of < 1.2, 1.1 to 1.4 by 2040-2069, and eventually 1.2 to 1.9 by 2070-2099. However, in the early autumn (Sep-Oct), the precipitation will decrease from present to 2040-2069 by a factor of approximately 0.9. In the "wet months" (Fig. 5b) there will be a very small increase in winter precipitation (factor of 1.1 or less), but a decrease in precipitation in May and June by factor of 0.9 to 0.82 (probably as a result of shift in spring to an earlier date and a shorter winter).

Ultimately, based on calibration bias, SDSM results were selected over k-nn results for further use in recharge estimation. It should be noted that at Grand Forks, the local climate is not modeled very well in the CGCM1 grid cell likely due to local convective precipitation and valley-mountain-rain-shadow effects, which have a strong influence on local precipitation. Notwithstanding the limitations of the downscaling methods at accurately predicting the observed climate, an important assumption is made that the GCMs can predict absolute changes in temperature and relative changes in precipitation, which then can be used to perturb current weather to arrive at future weather conditions.

CGCM1 downscaling was also used to predict basin-scale runoff for the Kettle River upstream of Grand Forks, as computed by Environment Canada [Whitfield and Cannon, 2000]. In future climate scenarios the hydrograph peak is shifted slightly to an earlier date, although the peak flow remains the same, and the summer baseflow period is extended and at lower levels [Scibek et al., submitted]. The relative importance of changes to the Kettle River discharge compared to recharge under future climate change scenarios is discussed later.

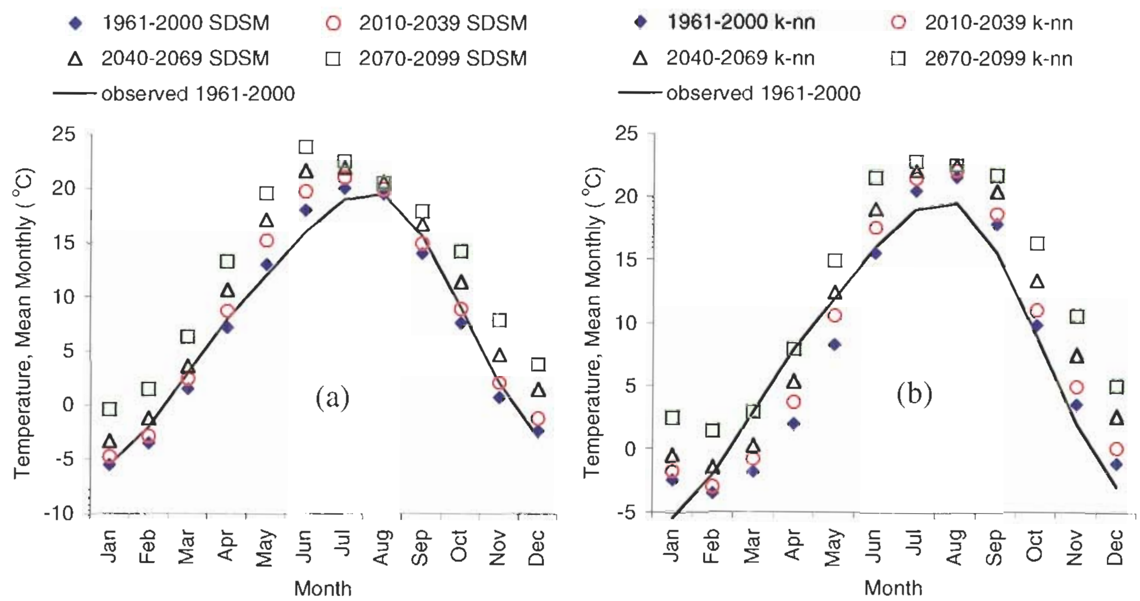

Figure (2) - 2: Man monthly temperature at Grand Forks, BC: observed and downscaled from CGCM1 model runs for current and future climate scenarios using a) SDSM and b) K-nn.

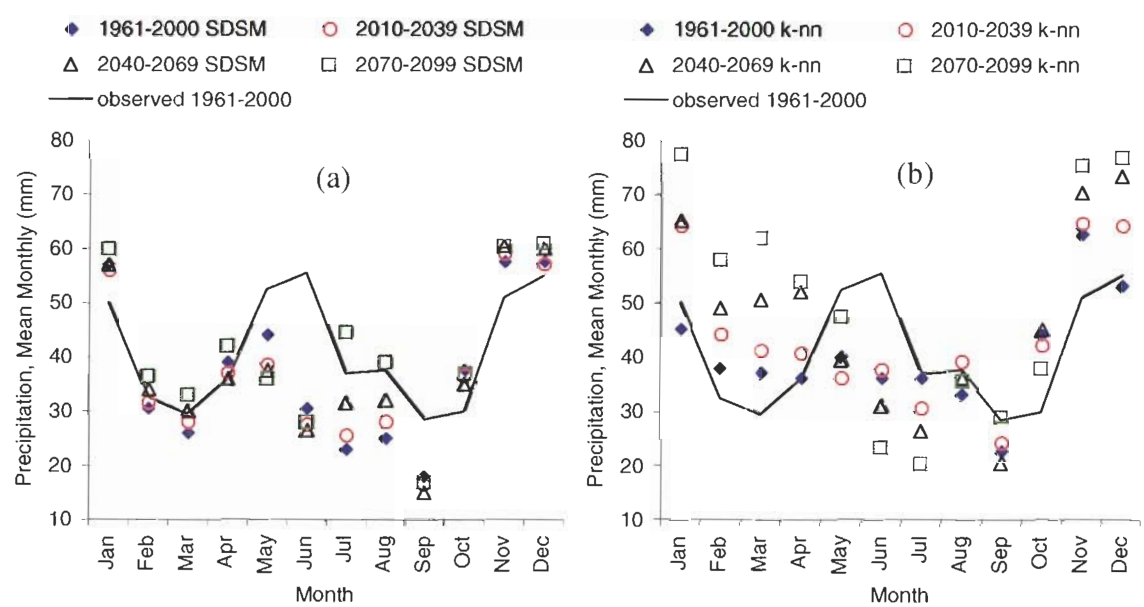

Figure (2) - 3: Mean monthly precipitation, at Grand Forks, BC: observed and downscaled from CGCM1 model runs for current and future climate scenarios using a) SDSM and b) K-nn.

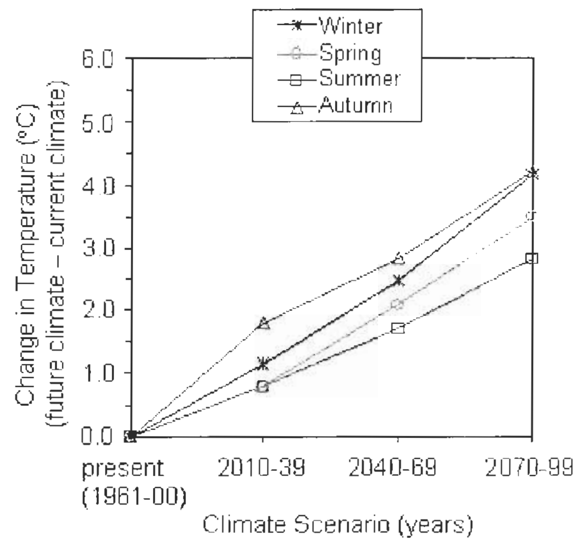

Figure (2) - 4: Absolute change in seasonal temperature predicted by CGCM1 model runs, after downscaling with SDSM for Grand Forks, BC.

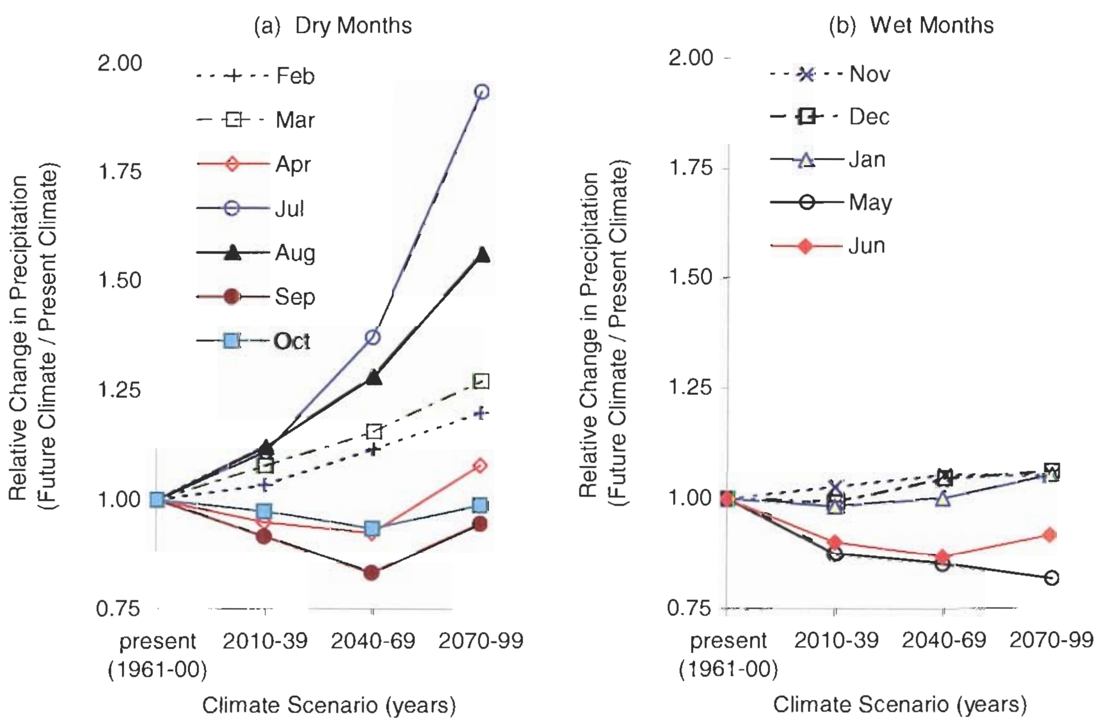

Figure (2) - 5: Relative change in monthly and seasonal precipitation predicted by CGCM1 model runs, after downscaling with SDSM for Grand Forks, BC. Comparing four seasons, and months within each season.

Daily Weather Inputs to the Recharge Model

The downscaled daily data already contain a stochastic component from SDSM downscaling [e.g., Diaz-Nieto and Wilby, 2005], but the poor downscaling results for precipitation did not allow us to use these data directly in a recharge model. Our approach was to compute change factors (relative and absolute), and redistribute them to daily time series using the stochastic series weather generator, LARS-WG [Racsko et al., 1991; Semenov et al., 1998]. LARS-WG utilizes semi-empirical distributions for the lengths of wet and dry day series, daily precipitation and daily solar radiation, and yields results that compare favourably to other weather generators, according to Wilks and Wilby [1999].

The base case is here defined as the average of the entire downscaled historical period, assuming that it is representative of pre-climate change conditions. Then, climate change scenarios are generated by perturbing the generated weather using the change factors to modify the base case – see Table 1. Each scenario consists of 100 years of generated weather (Fig. 6), noting that although generated weather runs of 1000 years converge better to specified "normals", there are diminishing returns of performance after 100 years. The length of generated weather time series is not meant to model actual changing climate year-to-year, but rather to model climate change step-wise for each scenario, and to generate a long enough weather time series to preserve and properly represent statistical properties for the site and the specified climate for the scenario.

Table (2) - 1: Climate scenario input (scenario file example) from SDSM to LARS-WG stochastic weather generator, for Grand Forks, BC. (Rain = precipitation relative change (future / base) or (base / base), mm/mm; Wet = Wet spell length relative change, mm/mm; Dry = Dry spell length relative change, mm/mm; T = temperature absolute change, °C; stdev T = standard deviation of temperature relative change, °C; SRad = solar radiation absolute change, MJ / m2 / day. Base case (1970-1999) for all months have Rain = 1.00, Wet = 1.00, Dry = 1.00, stdev T = 1.00, and T = 0.00 and SRad = 0.00.

base case (1970-1999)						
	Rain	Wet	Dry	T	sdev T	SRad
Jan	1.00	1.00	1.00	0.00	1.00	0.00
Feb	1.00	1.00	1.00	0.00	1.00	0.00
Mar	1.00	1.00	1.00	0.00	1.00	0.00
Apr	1.00	1.00	1.00	0.00	1.00	0.00
May	1.00	1.00	1.00	0.00	1.00	0.00
Jun	1.00	1.00	1.00	0.00	1.00	0.00
Jul	1.00	1.00	1.00	0.00	1.00	0.00
Aug	1.00	1.00	1.00	0.00	1.00	0.00
Sep	1.00	1.00	1.00	0.00	1.00	0.00
Oct	1.00	1.00	1.00	0.00	1.00	0.00
Nov	1.00	1.00	1.00	0.00	1.00	0.00
Dec	1.00	1.00	1.00	0.00	1.00	0.00

GF_CGCM1_2040_2069						
	Rain	Wet	Dry	T	stdev T	SRad
Jan	1.44	1.49	0.87	2.18	1.27	0.19
Feb	1.28	1.38	0.91	2.30	1.05	0.05
Mar	1.37	1.30	0.90	2.13	1.22	0.27
Apr	1.45	1.14	0.94	3.48	1.19	0.40
May	1.00	0.83	1.16	4.07	0.96	0.40
Jun	0.86	0.99	1.31	3.63	0.82	-0.19
Jul	0.73	1.00	1.15	1.90	0.66	0.26
Aug	1.11	1.21	0.96	1.10	0.71	-0.17
Sep	0.92	1.23	0.80	2.74	0.78	-0.44
Oct	1.03	1.01	1.28	3.80	0.97	0.25
Nov	1.15	1.39	1.44	4.03	0.84	0.05
Dec	1.39	1.11	0.60	3.88	0.78	-0.07

GF_CGCM1_2010_2039						
	Rain	Wet	Dry	T	stdev T	SRad
Jan	1.40	1.33	0.81	0.74	0.94	0.14
Feb	1.14	1.31	0.95	0.62	0.98	0.03
Mar	1.12	1.21	1.07	0.94	1.09	0.17
Apr	1.13	1.06	1.12	1.53	1.22	0.26
May	0.90	0.97	1.21	2.19	0.91	0.39
Jun	1.04	0.92	1.30	1.71	0.86	-0.08
Jul	0.83	1.10	1.01	1.00	0.78	0.15
Aug	1.19	1.27	0.95	0.41	0.85	-0.02
Sep	1.03	0.87	0.75	0.91	0.84	-0.24
Oct	0.97	1.17	1.24	1.27	0.94	0.11
Nov	1.04	1.17	1.29	1.33	0.84	0.00
Dec	1.21	1.02	0.82	1.15	0.92	-0.03

GF_CGCM1_2070_2099						
	Rain	Wet	Dry	T	stdev T	SRad
Jan	1.71	1.42	0.81	5.11	1.12	0.10
Feb	1.51	1.48	0.90	4.99	1.02	0.04
Mar	1.67	1.58	0.76	4.87	1.37	0.25
Apr	1.49	1.14	1.07	6.08	1.01	0.34
May	1.19	0.95	1.23	6.57	0.94	-0.31
Jun	0.65	0.84	1.37	5.82	0.65	-0.36
Jul	0.57	0.89	1.30	2.51	0.60	0.31
Aug	1.10	1.32	0.84	0.97	0.61	-0.56
Sep	1.27	1.53	0.71	3.93	0.55	-0.58
Oct	0.88	0.92	1.49	6.63	0.88	0.50
Nov	1.22	1.21	1.24	7.19	0.88	0.20
Dec	1.46	1.17	0.71	6.19	0.99	-0.12

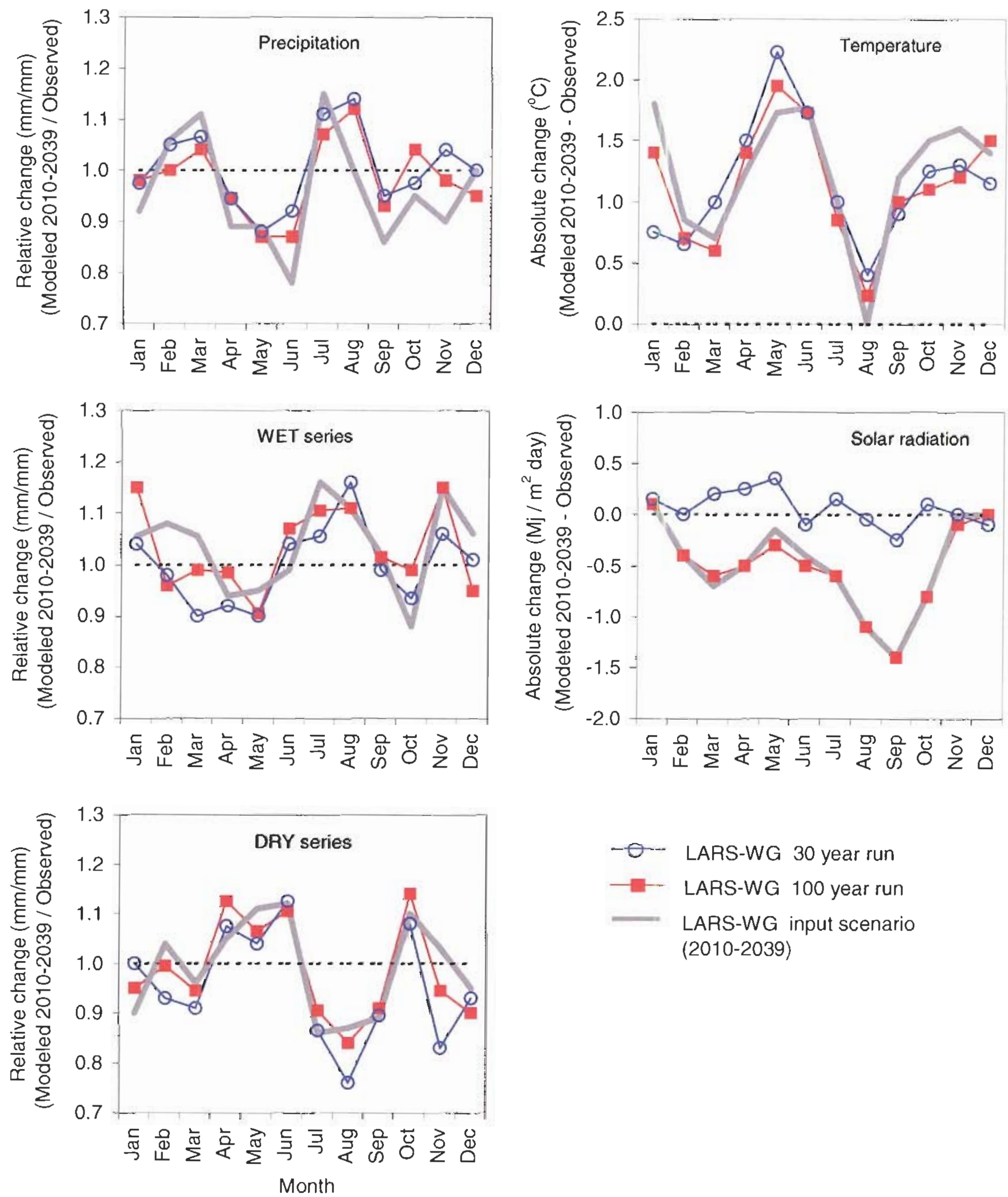

Figure (2) - 6: Comparing mean monthly weather parameters for the 2010-2039 climate scenario (the "inputs" to weather generator), with synthetic LARS-WG weather output, and comparing 30 and 100 year synthetic weather runs in LARS-WG: relative change in monthly precipitation, temperature, and solar radiation, expressed as either a ratio or absolute difference of monthly mean of LARS WG-generated value to the Observed value, where Observed represents the 1970-1999 historical monthly mean value.

Spatially-Distributed Recharge Modeling

In recharge modeling, GIS data-handling capabilities allow raster map algebra with classed maps of unsaturated zone properties, such as soil permeability (from soil type and land cover), to generate spatial distribution of recharge [Fayer et al., 1996]. There are many methods for recharge modeling [York et al., 2002], but the methodology presented here generates spatially-distributed and temporally-varying recharge zones, using a GIS linked to the one-dimensional US Environmental Protection Agency's Hydrologic Evaluation of Landfill Performance (HELP) model [Schroeder et al., 1994]. The program WHI UnSat Suite [Waterloo Hydrogeologic Inc., 2000], which includes the sub-code Visual HELP, is used to estimate recharge to the Grand Forks aquifer. HELP is a versatile quasi-two-dimensional layer model was originally designed for conducting water balance analyses and predicting hydrologic processes at landfills. However, in recent years it has been used effectively for estimating of groundwater recharge [Jyrkama et al, 2002], within the limitations of the model.

HELP uses numerical-solution techniques that account for the effects of surface storage, snowmelt, runoff, infiltration, evapotranspiration, vegetative growth, soil-moisture storage, and various engineering parameters (e.g., lateral subsurface drainage). The natural water-balance components that the program simulates include precipitation, interception of rainwater by leaves, evaporation by leaves, surface runoff, evaporation from soil, plant transpiration, snow accumulation and melting, and percolation of water through the soil profile. The profile structure can be multi-layered, consisting of a combination of natural (soil) and artificial materials (e.g., waste, geomembranes). In the current application of HELP, only natural geological materials consistent with those found in the Grand Forks aquifer were used. The rainfall-runoff processes in HELP are modeled using the US Department of Agriculture (USDA) Soil Conservation Service curve-number method [USDA 1986], and allows the user to adjust the runoff calculation to a variety of soil types and land-management practices. HELP uses different procedures to adjust the value of CN to surface slope, soil texture, and vegetation class. For purposes of simplicity, zero slope was assigned to each model layer. Although initial soil moisture can be specified, the code allows for values for the initial water-moisture storage of layers to be estimated, and then simulated over a one-year period. The values of moisture storage obtained from this simulation are then used as initial values, and the simulation starts again at year one.

The approach used is similar to that of Jyrkama et al. [2002], in which a methodology was developed for estimating temporally varying and physically based recharge using HELP for any MODFLOW grid cell. Our approach also depends on high resolution GIS maps (20 m grid) for defining recharge zones, and links these zones to MODFLOW model grids, although we developed a distinct methodology and code that links Visual MODFLOW v 3.1.84 [Waterloo Hydrogeologic Inc., 2004] to Arc GIS version 8.13 [ESRI, 2004] for input and output of MODFLOW simulations. Our method also differs from previous distributed-recharge methods in that we also estimate the distribution of vertical saturated hydraulic conductivity in the vadose zone and the thickness of the vadose zone, and incorporate irrigation return flow.

There are many physical properties of the subsurface that affect recharge to an unconfined aquifer and these have three-dimensional distribution; some change with time as well, such as soil moisture and depth to water table. The available data constrains the choice of some parameters through relatively good ground truthing, while other parameter values must be inferred from other information, and essentially estimated (Table 2). The parameters in Table 2 are listed in order of importance in each group. Usually, the type of local climate and, more specifically, seasonal distribution of precipitation will have dominant control on aerial recharge. The aquifer properties will control the actual amount of recharge into the aquifer, and are assumed constant in time, except unsaturated zone thickness, which will fluctuate seasonally.

Soil thickness was interpolated from 55 well lithologs (only 55 contained soil thickness data out of a total of 150) and dozens of soil pits. However, soils are expected to vary in thickness over micro-topography of the valley, thus any valley-wide interpolation of thickness would have very large error locally. With the exception of a few anomalous locations, the soil thickness is rather similar over the valley; the mean interpolated soil thickness is 0.92±0.21 m, and thus, the soil thickness was assumed to be simply 1.0 m in all percolation columns for recharge modeling (Fig. 7a).

Soil permeability maps were modified by land use to account for less permeable areas. Four representative permeability classes were created for very high, high, medium, and low permeability (see Table 3). Similar soils were combined into one category to reduce the number of categories to four, based on the spatial extent of each permeability class, to preserve the most representative soil types over the aquifer extent (Fig. 7b).

Table (2) - 2: Variables considered for recharge modeling.

Available variables	Estimated variables:
Climate:	
precipitation (daily to hourly),	evapotranspiration (daily)
	surface runoff (in low permeability soils)
Aquifer media properties:	
unsaturated zone thickness (depth to water table)	unsaturated zone hydraulic properties from lithology at point locations (equivalent saturated hydraulic conductivity)
soil type (permeability)	soil thickness distribution (assume uniform 1 m due to lack of data of adequate spatial resolution)
Ground surface properties:	
vegetation cover	effect of vegetation on recharge
irrigation rates and areas affected	return flow to recharge
elevation and slope of ground surface (valley floor topography)	

Table (2) - 3: Permeability classes for soils in Grand Forks aquifer area.

Vertical percolation layer in HELP	Vertical Kz_{sat}		Permeability
	(cm/s)	(m/d)	
Silty Loam	1.90E-04	0.164	low
Loam	3.70E-04	0.320	
Fine Sandy Loam	5.20E-04	0.449	
Sandy Loam	7.20E-04	0.622	moderate
Loamy Fine Sand	1.00E-03	0.864	
Loamy Sand	1.70E-03	1.469	high
Sandy Gravelly Soils	5.80E-03	5.011	v. high

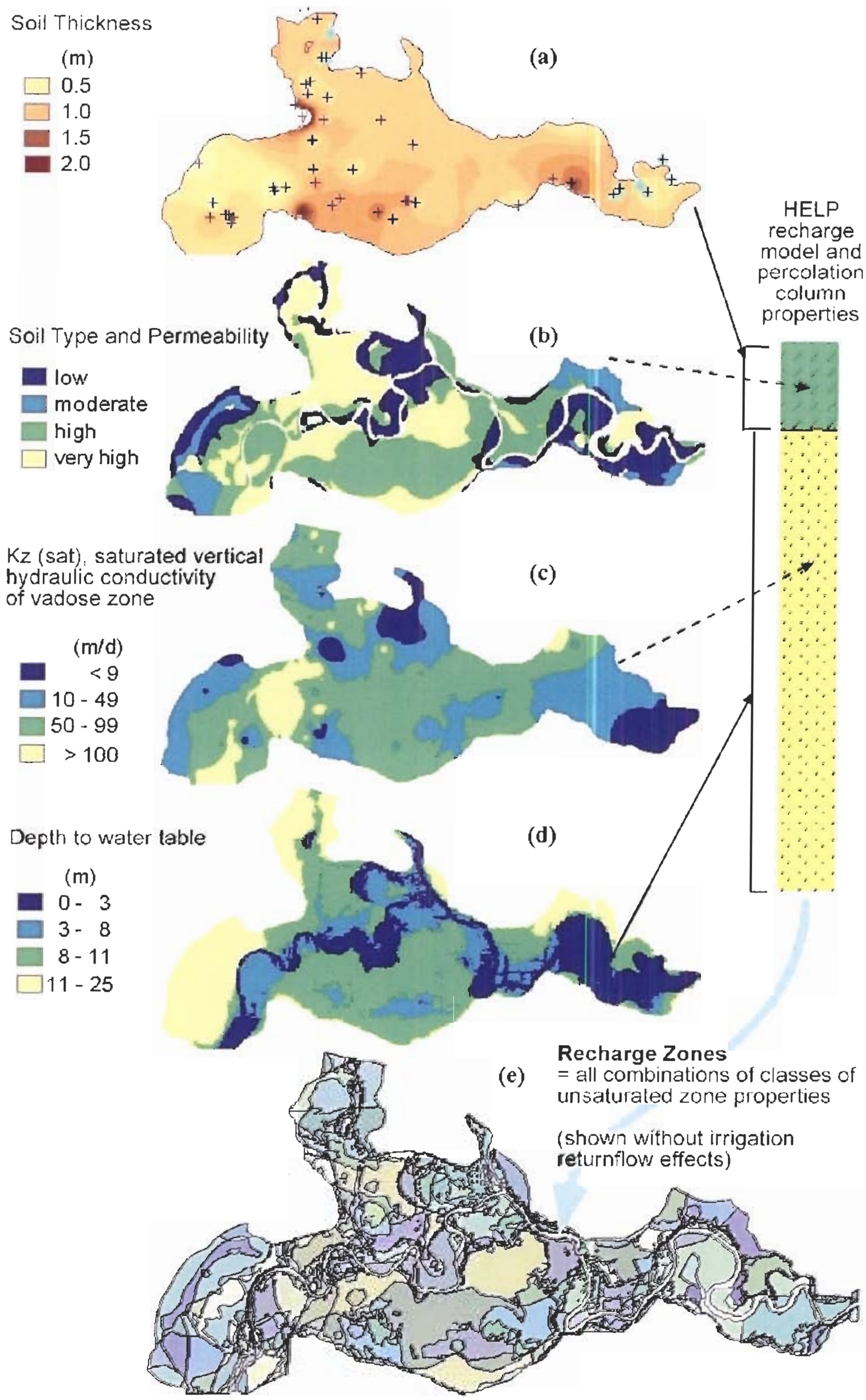

Figure (2) - 7: (a) Soil thickness, (b) re-classed Kz map of unsaturated zone above water table in Grand Forks aquifer, (c) soil permeability classes (see Table 3), (d) depth to water table classes, (e) resulting recharge zones.

Saturated hydraulic conductivity estimates were estimated for geologic units encountered above the water table. Well lithology data were standardized and classified using a custom code in order to simplify the data. Up to three material descriptions were retained for each depth interval. Saturated (assumed vertical) hydraulic conductivity (Kz), specific storage (Ss) and specific yield (Sy) for each material type were assigned based on representative values in the published literature, and were constrained by parameters estimated from pumping test data. Geometric means of the Kz values were calculated for each layer in each well where more than one material type was recorded. A manual examination of the output data was carried out in order to ensure that the calculated hydraulic conductivities were consistent with the dominant sediment description in the original well log. In only a few cases (<10) were modifications made as a result of the standardization scheme not correctly identifying the dominant material types. Equivalent Kz was computed for each well point location, assuming homogeneous and isotropic "units". Kz values in 285 wells ranged from a maximum of 1000 m/day to a minimum of $1x10^{-6}$ m/day, median value of 13 m/day, and quartile values of 100 and 0.14 m/day. The Kz values in the vadose zone were interpolated using Inverse Distance Weighed interpolator, and computed on representative vertically averaged log Kz values at all available point locations where lithologs exist. After interpolation, the inverse logarithm (i.e., 10^[Log Kz]) of the interpolated raster was computed, and converted to units of m/day. Four classes were chosen as $1x10^{-6}$ to 0.14, 0.14 to 13 m/day, 13 to 100 m/day, and 100 to 1000 m/day (Fig. 7c). The representative Kz values for each material in the HELP soil columns were 315, 40, 1.4, and 0.015 m/day (mid value in each class).

Depth to water table was estimated from the difference between ground surface and a numerically-derived static groundwater table [Allen et al., 2003]. This assumption is reasonable as the depth to the water table is usually much larger than the variation in groundwater level, except in the low-lying river floodplain region where river effects dominate the water levels, not recharge [Scibek et al., submitted]. Depths to the water table in 285 wells ranged from 1.5 m to 46.8 m, with a median of 10.1 m, and quartile values of 6.1 m and 12.9 m. The depth classes were based on quartiles of distribution: 0 to 6 m, 6.1 to 10 m, 10.1 to 12.9 m, 13.0 to 47.0 m, with roughly 25% of aquifer area in each of four categories (Fig. 7d). Representative sediment columns in HELP were assigned representative mid-class depths: 3, 8, 11, and 25 m.

Recharge zones were defined for a 50 m raster grid through cross-classification of maps of all important variable distributions (Fig. 7e), resulting in 65 zones (zone 1 was the default zone in MODFLOW with no recharge). A recharge zone is any unique combination of soil permeability class, hydraulic conductivity class, and depth to water table class, thus, the number of combinations of soil permeability, Kz, and water depth, was (4 x 4 x 4 =) 64 soil columns. As stated earlier, soil thickness was assumed the same for all columns. There is a degree of uncertainty in each of these properties because data come from various sources and formats. In this study, the limiting variable is soil type (originally soil polygons), and the most uncertain is Kz as representative values were estimated, whereas depth to water table could be represented at 20 m grid or smaller with reasonable accuracy. Higher spatial resolution would have required more sub-classes in each variable and resulted in many more recharge zones; however, more combinations were used in sensitivity analyses of HELP model performance.

Irrigation Return Flow

A better representation of recharge in an agricultural area takes into account the amount of water that is returned to the aquifer when the land is irrigated. This is commonly referred to as irrigation return flow (from the aquifer perspective). In all pumping model scenarios, the recharge zones were modified by superimposing estimated irrigation return flow to the aquifer. Generalized estimates of return flow were obtained through consultation with experts in irrigation practices (roughly 25% of the amount of irrigation for the types of crops present). We assumed constant irrigation return flow in irrigated fields for each month (June to August only) in all present and future climate scenarios. As future irrigation predictions would require estimates of population change, land use change, technology change, and climate change and associated feedbacks, we were unable to predict changes in irrigation demand, and thus, changes in future irrigation return flow. Nonetheless, this is certainly an aspect that should be considered if such information could be predicted. In the MODFLOW model, additional recharge zones were created to represent the modified recharge after addition of return flow from irrigation.

The final step involved transferring recharge values into the transient groundwater flow model. Each MODFLOW cell had an independent schedule for recharge. Evapotranspiration was taken into consideration in HELP model from weather inputs, and precipitation was assumed to be uniform over the aquifer in this valley. Custom codes were written to update MODFLOW files with recharge "zones" and schedules.

Sensitivity of HELP to Model Inputs

The performance of the HELP model is adequate compared to other similar models for most conditions [Scanlon et al., 2002]; however, it is important to evaluate the sensitivity of modeled recharge to HELP input parameters. The HELP model results showed a very small effect (< 5% change) for: type of stand of grass, wilting point, field capacity, and initial moisture content (results not shown). A moderate effect was found in soil thickness (Fig. 8e) and porosity of percolation layer (Fig. 8f). As soil thickness increased, the modeled recharge decreased, but only very strongly from April to June (wet months at this site) as this effect is precipitation- and temperature-dependent. The strongest effect on HELP model recharge results was variation of the depth of the vadose zone or percolation layer (Fig. 8c-d). Similar effects, but with different magnitudes, were observed for different soil types or permeability of soil (Fig 8b) and for saturated vertical hydraulic conductivity of vadose zone (Fig 8a). The effects were seasonal and most pronounced in spring to early summer, again due to combination of precipitation and temperatures that control evapotranspiration and infiltration rates (together with unsaturated zone properties).

The high sensitivity of recharge models such as HELP to unsaturated zone properties suggests that spatial distribution of such properties must be accounted for in recharge modeling for climate change impacts assessment in surficial aquifers.

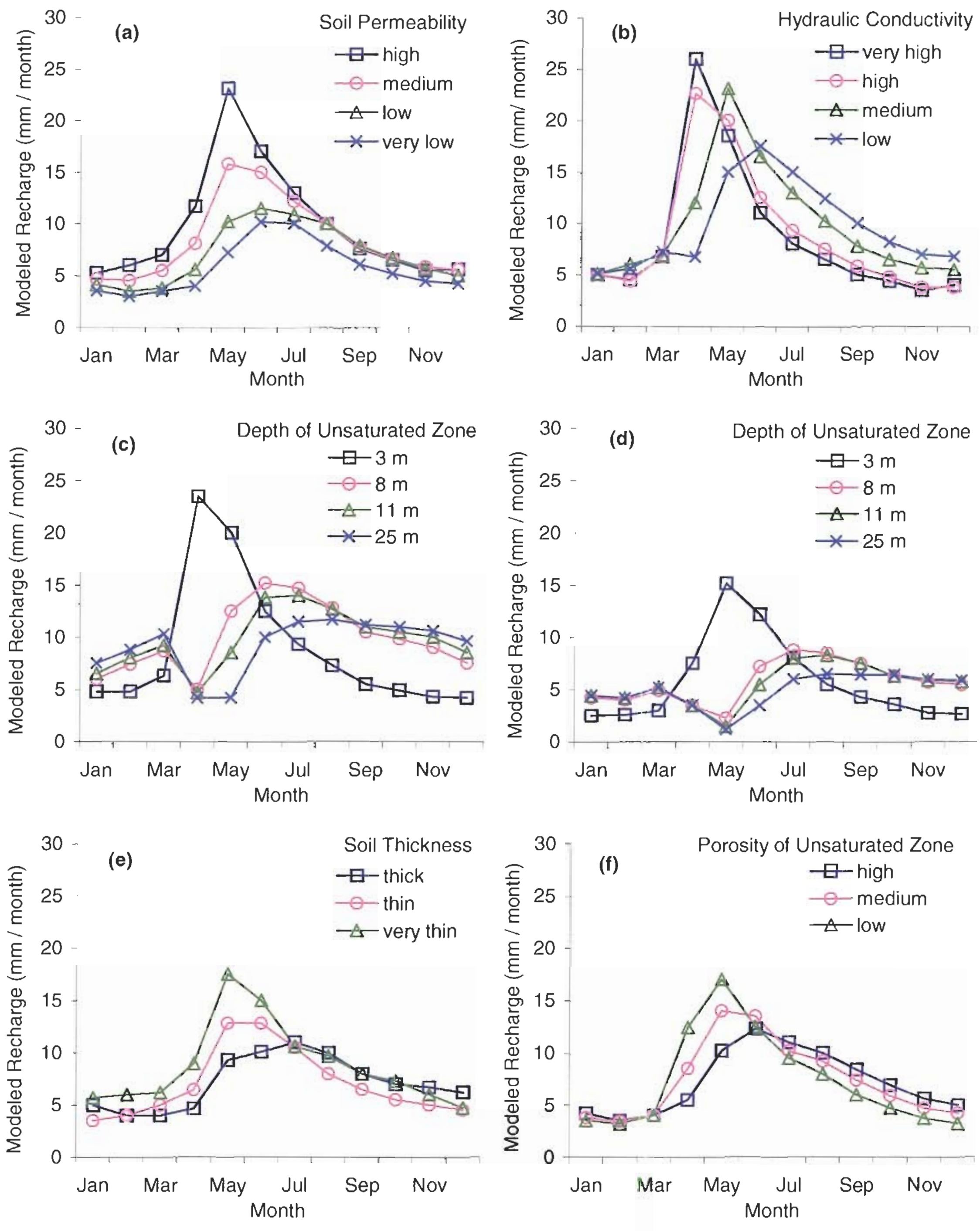

Figure (2) - 8: Sensitivity of recharge estimates modeled with HELP to (a) saturated vertical hydraulic conductivity of vadose zone, (b) soil permeability, (c – d) depth of vadose zone and soil permeability, (e) soil thickness, (f) porosity of vadose zone material.

Historical Recharge Results

Previous recharge modeling [Allen et al., 2003] used a uniform annual recharge value for the Grand Forks aquifer of 135.5 mm/year, or approximately 28% of precipitation. According to this study, mean annual recharge varies considerably across the 64 recharge zones (Fig. 9), ranging from near 30 to 120 mm/year (10% and 30% of mean annual precipitation). The low recharge values were associated with areas of thick gravels in the terraces, which can absorb a large amount of rainfall before reaching saturation, and because this region is relatively dry, the modeled recharge is low in areas with large depth to water table. It would be expected that infiltrated water below the root zone would continue to water table, and after some lag time, contribute to recharge, so this result is surprising. The initial moisture contents of these deep layers are unknown, but were estimated in HELP by first running one year of weather, and using the moisture content at the end of that year as the initial value. Nonetheless, the recharge model was not greatly sensitive to initial moisture content in a one year simulation conducted at daily time steps.

Mean monthly recharge (to the inset area shown in Fig. 9) is shown in Figure 10. Recharge follows the annual distribution of precipitation, when summer rainstorms supply most intense rainfall and most of recharge to aquifer is from rainfall. Note that the range in percentages of mean annual precipitation is typically smaller than range in percentages of monthly precipitation, due to seasonal variation in precipitation and averaging on annual time scales. Monthly recharge varies from <2 mm/month to >12 mm/month, or between 10% to 80% of monthly precipitation (Fig. 10). Most of the recharge is received in spring and summer seasons, while in winter the ground is frozen and snow melt does not occur. The autumn season is relatively dry. In spring time, by monthly value, the aquifer receives 40% to 80% recharge from precipitation, depending on soil properties and aquifer media properties, while in the summer, the values are 30% to 50%. During late summer the aquifer receives 60% to 90% of precipitation, but the overall recharge amount is small because rainstorms are infrequent. The LARS-WG preserves the intensities of rain events; we observe that if a high intensity event occurs during the late summer (such as a thunderstorm), it rains heavily and most of the water infiltrates the aquifer. If it were to rain slowly and over a longer time, much more of it would evaporate. This type of relation may be very different in other climate regions and in other aquifers where high intensity rainfall events may lead to increased runoff and less infiltration.

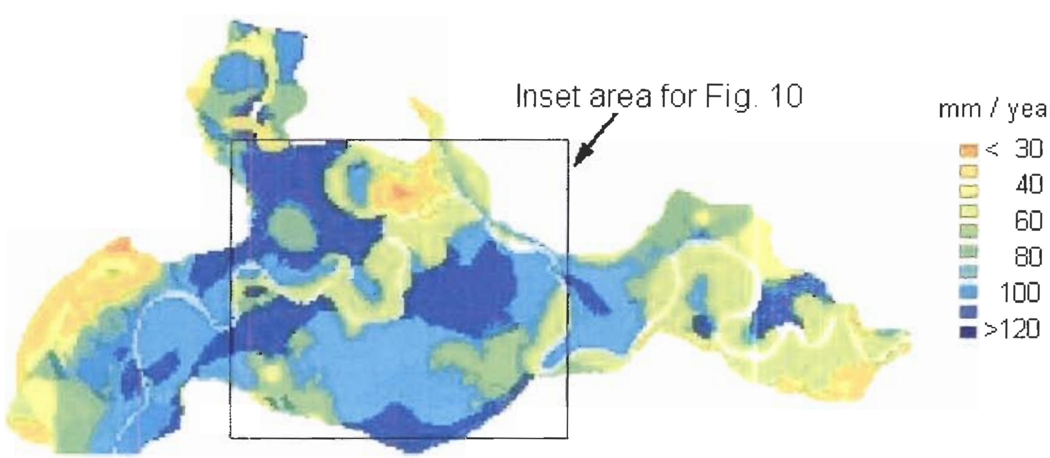

Figure (2) - 9: Historical mean annual recharge to the Grand Forks aquifer for the historical climate scenario (1961-1999), modeled in HELP and assigned to recharge zones.

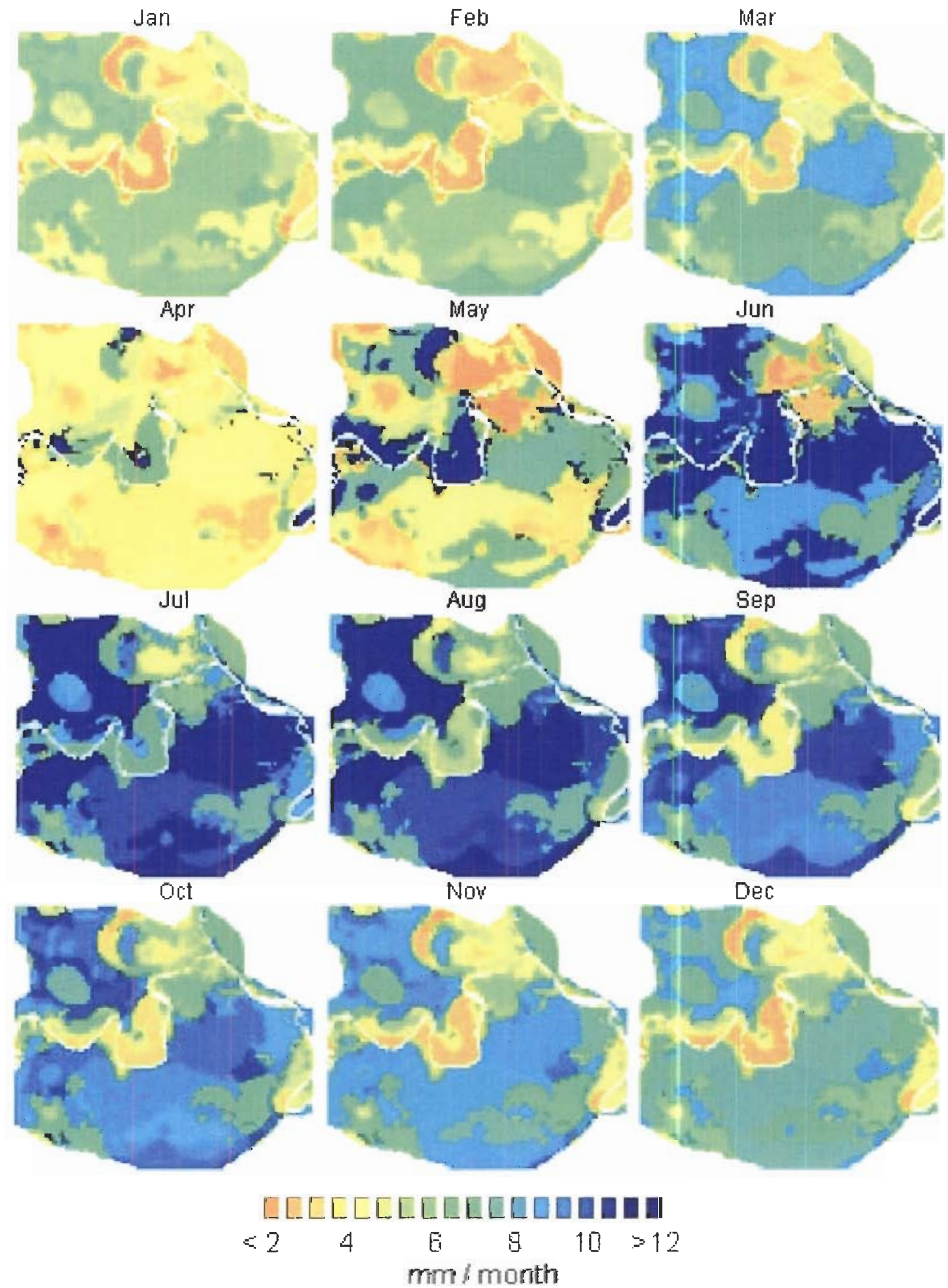

Figure (2) - 10: Historical mean monthly recharge maps for inset area (central portion of valley) as shown in Figure 9.

Predicted Recharge Changes for Future Climates

The predicted changes in mean annual recharge were converted to percentage differences: (future – historical) / historical. The 2010-2039 climate scenario has a predicted 2 to 7 % increase from historical mean annual recharge and there are no predicted decreases (Fig. 11a). Monthly recharge results (not shown) have the lowest recharge occurring in January through May, the highest recharge occurring in June to September, and October through December receiving moderate recharge.

The 2040-2069 climate scenario has a predicted 11 to 25 % increase from historical mean annual recharge, also without any predicted decreases (Fig. 11b). Monthly recharge results (not shown) have the lowest recharge occurring in January through May, the highest recharge occurring in June to September, and October through December receiving moderate recharge.

Figure 12 shows the percentage of precipitation that contributes to recharge for the most commonly-occurring recharge zones in the aquifer, for the three different climate scenarios. Recharge as percentage of precipitation increases in future climates, but in 2040-2069, some zones receive less recharge as a percentage of annual precipitation compared to 2010-2039 and present climates.

We did not specifically model the potential changes to snowmelt timing and resulting earlier spring thaw in the soil, and also a potentially longer growing season. However, by the time the spring rains come at Grand Forks the ground is usually thawed and rain on snow is not significant. Changes in the amount of snow on the ground prior to snowmelt, caused by climate modification, were not modeled as there is no reliable information suggesting any such changes and in which direction these might occur.

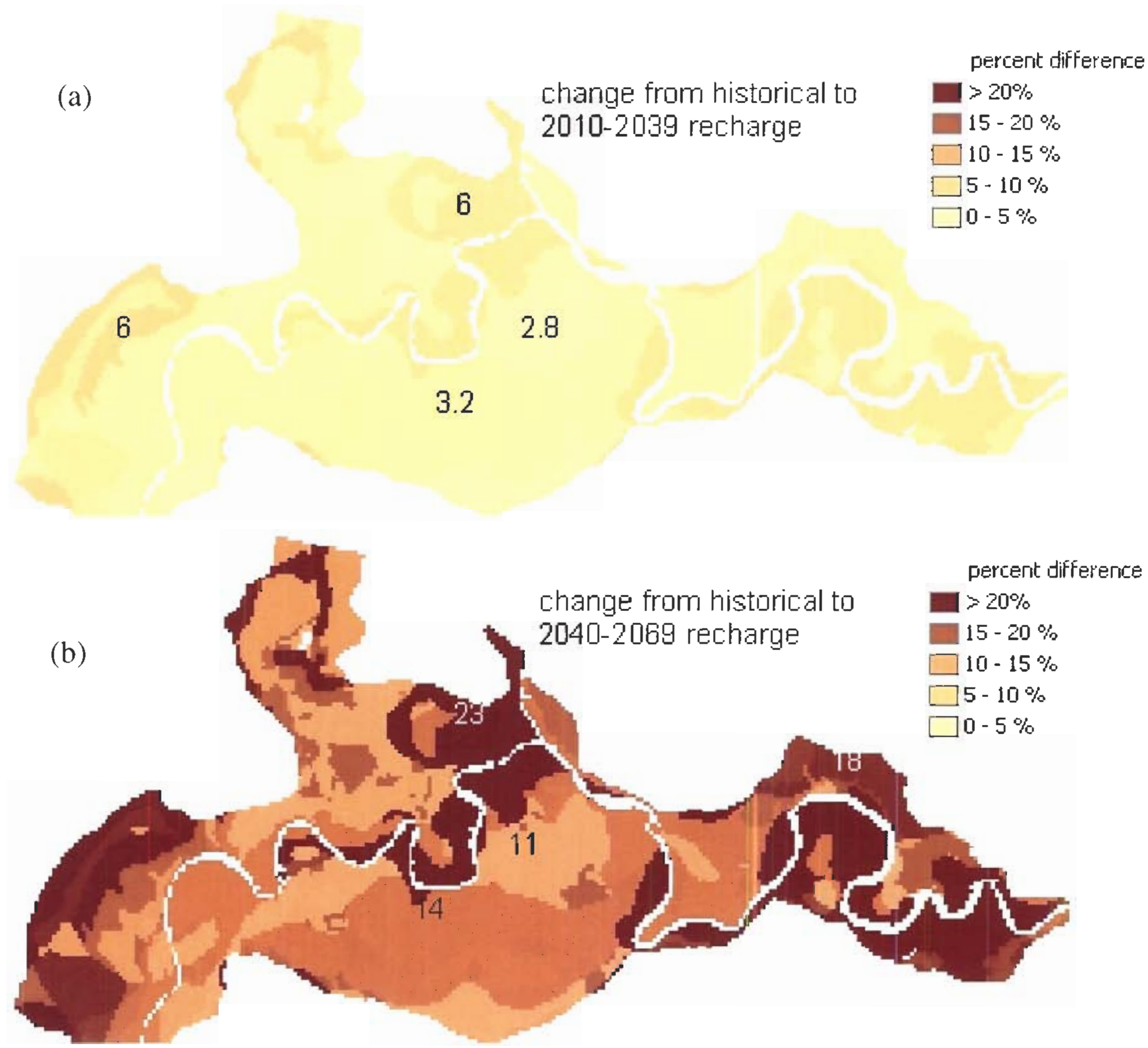

Figure (2) - 11: Percent change in mean annual recharge to the Grand Forks aquifer modeled in HELP and assigned to recharge zones: between (a) 2010-2039 and historical, (b) 2040-2069 and historical. Historical climate scenario (1961-1999).

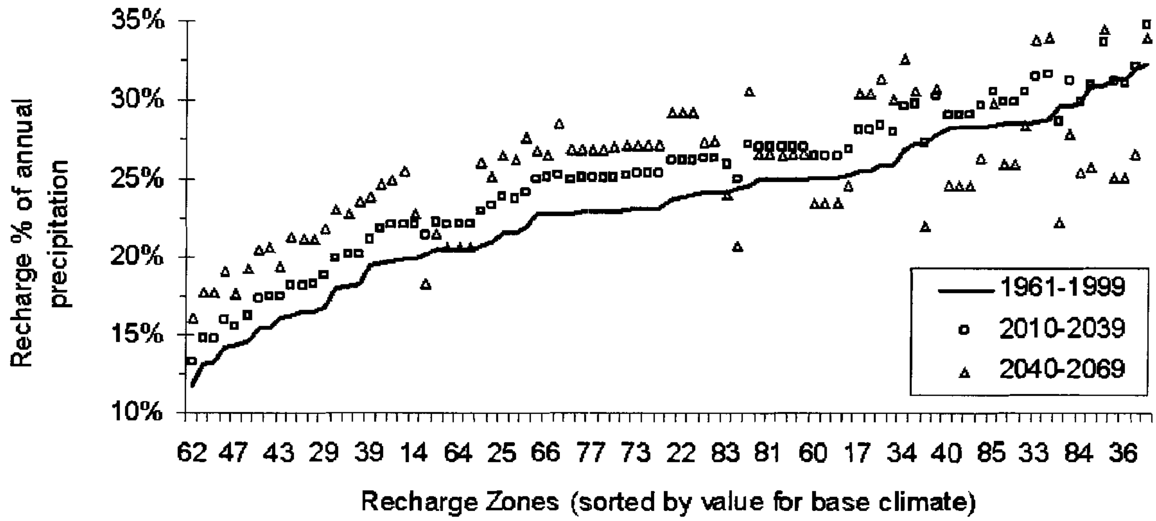

Figure (2) - 12: Recharge as percentage of annual precipitation for most common recharge zones in the Grand Forks aquifer for three climate scenarios.

Aquifer Model Development

The numerical groundwater flow model was developed using standard modelling methodology, and implemented in Visual MODLOW [Waterloo Hydrogeologic Inc., 2004]. The model domain (active cells) consists of unconsolidated sediments infilling a deep valley eroded into metamorphic rocks. The valley attains a maximum depth of approximately 250 m below ground surface, based on surface modeling from all outcrops and extrapolated valley sides, but typical sediment thickness is about 50 to 100 m. The stratigraphic sequences in the Grand Forks valley are poorly understood, particularly at depth. However, based on the low permeability nature of these deeper sediments, the lack of deep information is not anticipated to affect the results for the upper aquifer horizons. Previous interpretations of the 150 well lithology logs assumed a uniformly layered paradigm of hydrostratigraphy [Wei et al., 1994]. The unconsolidated sediments thicken toward the middle of the valley, have presumed horizontal stratigraphy in a layered model, and the topmost coarse grained sediments form the Grand Forks aquifer. Re-interpretation of the valley hydrostratigraphy was aided by considering other similar valleys in BC, where the basal units are commonly silt, clay and gravel, overlain by thick glaciolacustrine silts [Fulton and Smith, 1978; Fulton, 1984; Clague, 1981, Ryder et al., 1991], and capped by Holocene sandy and gravelly outwash and floodplain deposits and paraglacial alluvial fans. The hydrostratigraphic units in this model (Fig. 13) were modeled manually, with the aid of GMS version 4.0 [Brigham Young University, 2002], in three-dimensions from standardized, reclassified, and interpreted well borehole lithologs [Allen et al., 2004]. Solid models, representing different hydrostratigraphic units, were constructed and converted to 5 layers in MODFLOW, as is typically done with complex multi-layer aquifer systems [Herzog et al., 2003].

In this paper, we describe the results of the homogenous aquifer properties model. The homogeneous case was selected because it is simple, errors in the hydraulic conductivity data are reduced by averaging over the aquifer area, and the least assumptions are made about local geology. Homogenous and isotropic values of hydraulic conductivity were assigned to each layer, based on average values determined from pump test data. The model included the respective historical or future predicted river hydrograph boundary condition [Scibek et al. submitted]. The steady state model was calibrated to historic static water levels in over 200 wells, while the transient model was verified against one high quality observation well with monthly records and overall anticipated transient behaviour of the aquifer. A detailed model description is beyond the scope of this paper, but can be found elsewhere [Allen et al., 2004].

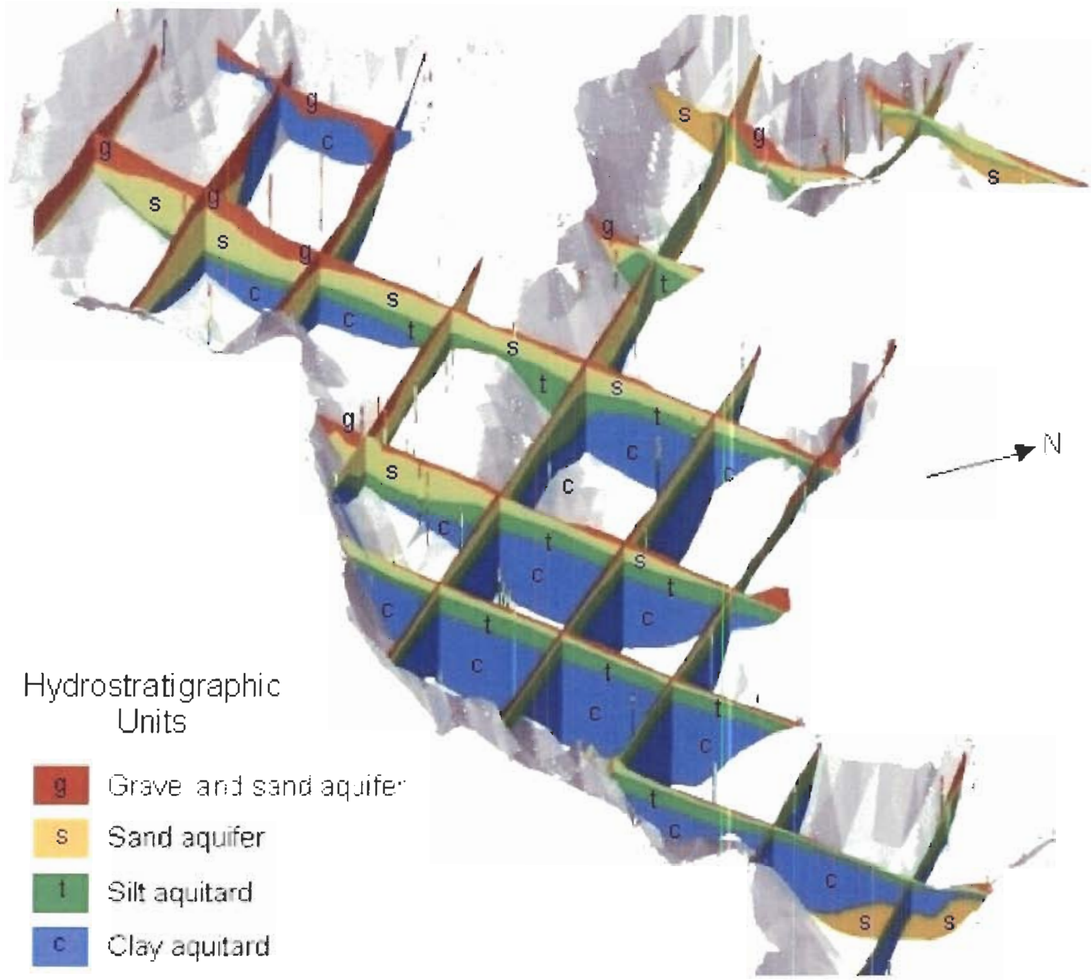

Figure (2) - 13: Fence diagram of hydrostratigraphic units in the Grand Forks valley.

Climate simulations were also carried out for a heterogeneous model, whereby the upper aquifer layer was represented not by a uniform K value, but rather a heterogeneous distribution of K values. The effects of aquifer heterogeneity on climate change impacts modeling are discussed by Allen et al. [2004]; however, we note that valley-wide results are entirely consistent with those presented here in that the same general trends are observed, except that in the heterogeneous model, there is more local variation in the predicted changes in water level.

Spatially-distributed recharge was mapped for each climate scenario and applied to the numerical groundwater flow model. Various simulations were undertaken to test different boundary conditions, for example, different recharge scenarios (historical or future predicted), pumping versus non-pumping, or type of recharge distribution over the aquifer area (uniform or spatially-distributed).

The PCG2 (Generalized Preconditioned Conjugate-Gradient Package) had the most success in converging on solutions, and was used for all simulations. The solver was used with modified incomplete Cholesky preconditioning. Cell re-wetting resulted in particular challenges. Wetting threshold was set equal to the precision of solver in head convergence, to be consistent. This would cause re-wetting at every wetting interval. The wetting interval was large to allow the model to non-rewet automatically during most iterations, in order to allow convergence. During cell re-wetting, the model often ran over 100 to 200 outer iterations without convergence, thus the maximum number of outer iterations were set to 500 (normally set at 25). Similarly, the number of inner iterations was increased to 100 (from typical 10) because of problems with convergence and during re-wetting of dry cells. Re-wetting was eventually done manually, by lowering the head convergence criterion in the solver during run-time to very small number. In effect this involved setting >30 outer iterations in model solution to stabilize the heads prior to cell re-wetting, then re-wetting would occur at iteration 30, and the solver head convergence criterion was changed back to 0.001 value, after which the model would converge at that time step and continue solving. Re-wetting was specified during stress-periods as follows (mostly during rise in river hydrographs in spring time, and always at first stress period, and always during recovery from pumping in pumping models). During falling water levels in rivers, cell re-wetting was not found to be a problem, and typically the model would perform 10 to 20 outer iterations per time step, with inner iterations decreasing during convergence. However, the solver would take 100 to 80 inner iterations per outer iteration during periods of fast changes in river stage and dry-cell re-wetting.

Model calibration incorporated approximately 300 observation wells where static (historic) water levels were available. In addition, the transient response of a provincial monitoring well was used. The normalized RMS (root mean squared) error for residuals between calculated and observed head was roughly 8%, and most data fall within the 95% confidence interval.

Finally, a water balance for each model was computed using Zone Budget (ZBUD) in MODFLOW [Harbaugh, 1988]. The zones represent different irrigation districts within the unconfined aquifer, the river floodplain, and deeper model layers. Temporal changes in mass balance components were graphed to show relations between pumping, storage, recharge, and flow for each climate scenario.

Sensitivity to Recharge Distribution

In this section we discuss the sensitivity of the groundwater flow model results to method of representation of recharge. Two sensitivity scenarios were run: 1) mean annual recharge as the recharge input, where the values are spatially distributed among the recharge zones (i.e., there is spatial variation in mean annual recharge) (scenario 5A); and 2) temporally variable recharge rates with uniform spatial distribution (one recharge zone) (scenario 5B). The single recharge zone in 5B was chosen arbitrarily to represent "high" recharge, or a shallow depth to water table and high hydraulic conductivity of the unsaturated zone. In most areas of the aquifer the variation in depth to water table and soil permeability class would produce a range of recharge zones, from high to low recharge values as modeled in HELP. The one "high recharge" zone is expected to have a higher aquifer recharge compared to scenarios with multiple recharge zones. Each scenario (5A and 5B) is compared to the historical base case (scenario 1A) using spatially and temporally-varying recharge. Only the non-pumping historical climate scenarios were used to test the sensitivity of the model to recharge distribution. The two scenarios are compared directly using water level difference maps; the effects of simulated future climate change are difficult to observe on head distribution maps because the high hydraulic gradient in the Grand Forks valley dominates all other trends. Head difference maps subtract gradients and pumping effects (constant pumping rates) and show only effects of modification of model inputs on modeled groundwater levels.

The impact of spatial distribution of recharge on water levels is much greater than that of temporal variation in recharge (Fig. 14) Recharge zonation reduces recharge from a uniform "high" value to a range of values depending on recharge zone, thus differences between scenario 5B and scenario 1A reflect the advantage of considering spatially-dependent soil permeability and water table depth (Fig. 14). In the Grand Forks aquifer, the model is sensitive to recharge only away from river floodplain, and the maximum change expected in water table elevation is between 10 and 50 cm, but typically about 20 cm. For example, at days 101, 131 and 265 (Fig. 14), the difference in water levels (between scenarios 5B and 1A) away from the river is less than 10 cm. In some portions of the aquifer, such as the floodplain area where river levels control groundwater levels

[Scibek et al., submitted; Allen et al., 2004], the model is not sensitive to recharge representation.

The analysis of flow components (Fig 15) also suggests the greater importance of spatial distribution of recharge representation in the model (compared to one recharge zone) over the monthly distribution of recharge representation (compared to mean annual). If the mean annual recharge value is applied to the model, as would be the case when climate data are lacking or when temporal recharge estimates are unavailable, then modeled water levels are within 10 cm of that modeled with temporally variable recharge.

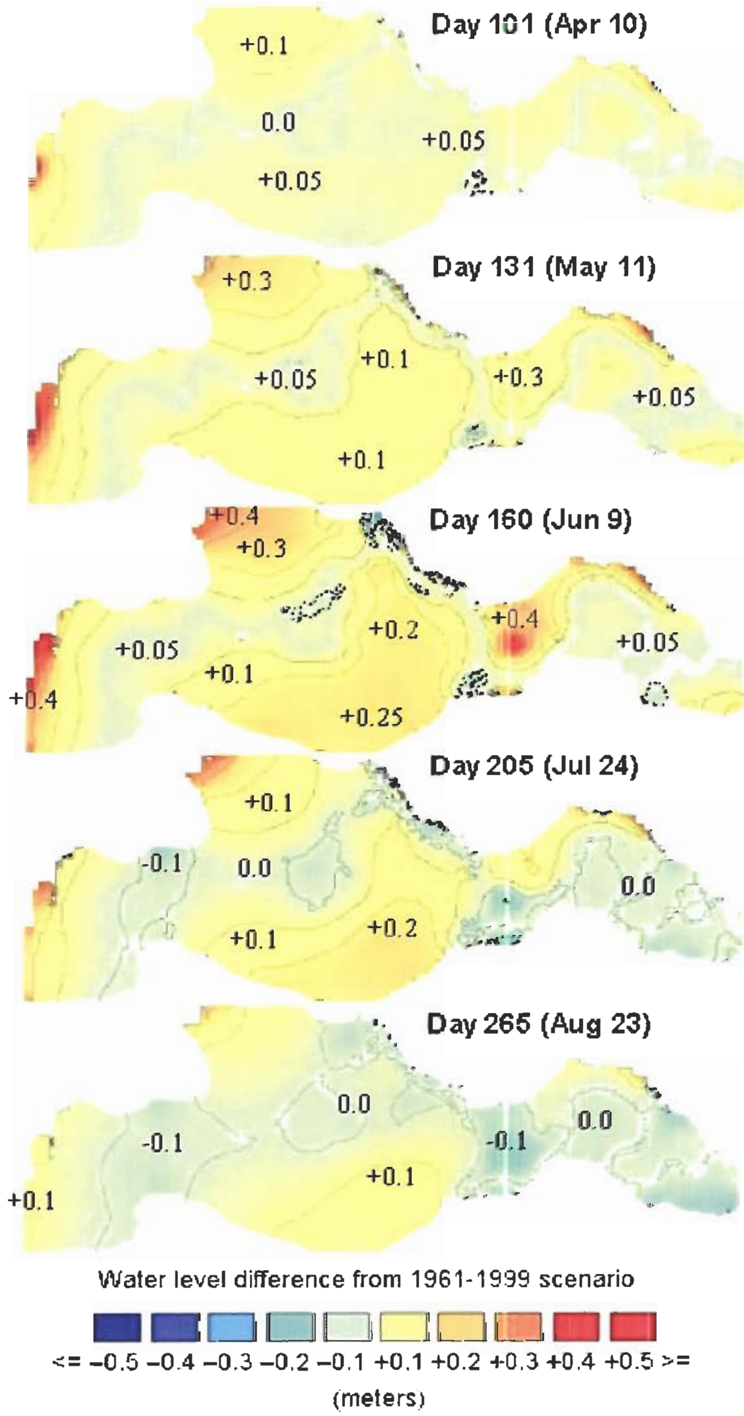

Figure (2) - 14: Water level differences (measured as head in layer 2 of unconfined aquifer) calculated from the difference between model scenario outputs (5B - 1A). Scenario 5B represents spatially non-distributed recharge (i.e., mean monthly recharge applied to a single zone). Scenario 1A is the historical base case using spatially and temporally-varying recharge. Maps by time step in days 101 to 265. Contours shown at 0.1 m interval. Zero contour is dashed line.

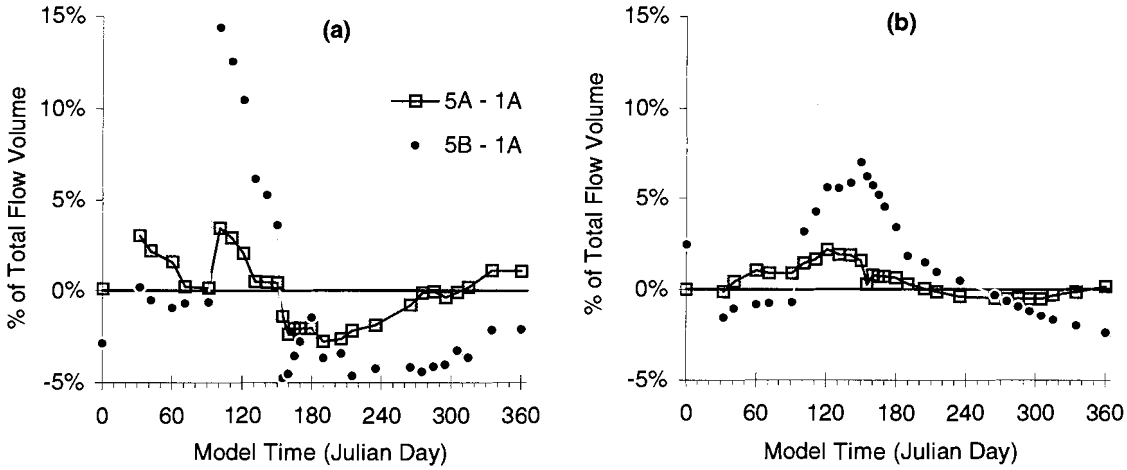

Figure (2) - 15: Effect of both spatial and temporal distribution of recharge on (a) storage and (b) flow from/to other zones, in non-pumping groundwater flow models. Values are calculated from the difference between model scenario outputs (5A - 1A) and (5B – 1A), where 1A is historical base case using spatially and temporally-variable recharge; 5A is temporally-constant recharge (i.e., annual recharge applied in a distribute fashion); and 5B is spatially-non-distributed recharge (i.e., mean monthly recharge applied to a single zone). % calculated from (OUT – IN) / average (OUT + IN).

Model Results

The groundwater flow modeling scenarios were based on: 1) climate scenario and corresponding predicted recharge and predicted Kettle River hydrograph, 2) pumping or no pumping, 3) type of recharge distribution over the aquifer area (uniform or spatially-distributed), 4) type of aquifer representation (homogeneous or heterogeneous K distribution).

Based on groundwater flow model simulation results for historical climate the volumetric recharge accounts for only 1 to 7% of other flow components, such as flow between zones and storage. Figure 16 graphs recharge for zones 4 and 5, which represent the large Sion (2 zones) and Big Y (1 zone) irrigation districts in the valley. Aside from seasonal precipitation trends, many zones also have bi-modal distribution, with a smaller peak of recharge in late winter, which corresponds to snowmelt in the valley. The inter-zonal differences are due to soil surface and subsurface properties and their hydraulic properties, as calculated by recharge model, also including differences in surface runoff.

Irrigation return flow increases the recharge by 10 to 20% in most zones (e.g. zones 4 and 5 on Fig. 16). The importance of return flow in ZBUD zones depends on % irritated area. Recharge flow volumes are small for such a large zone compared to other flow terms (recharge is 2% of other flow rates, such as flow to/from other zones for this zone). In the late time steps of the model year, the recharge rates for the pumping model are higher than for non-pumping model, possibly as a result of drawdown in some areas. Drawdown creates more "dry" cells in overlying aquifer layers in the MODFLOW model, and redirects more recharge to the silt layer below.

Kettle River Water Levels

High quality monthly water level records in an observation well located in the river floodplain indicate that the groundwater levels fluctuate predictably and regularly with rising and falling Kettle River stage over each annual hydrologic cycle [Allen et al., 2004]. The river carries an order of magnitude more flow (per unit time) than exchanges between the river and the aquifer (i.e., a maximum 41 m^3/s, which translates to 15% of river flow during spring freshet), so effects of groundwater inflow on river discharge and stage are presumed negligible. In areas distal from the river, the effect is relatively small, but significant, and it varies over the year. A detailed water balance has been calculated [Scibek et al., submitted].

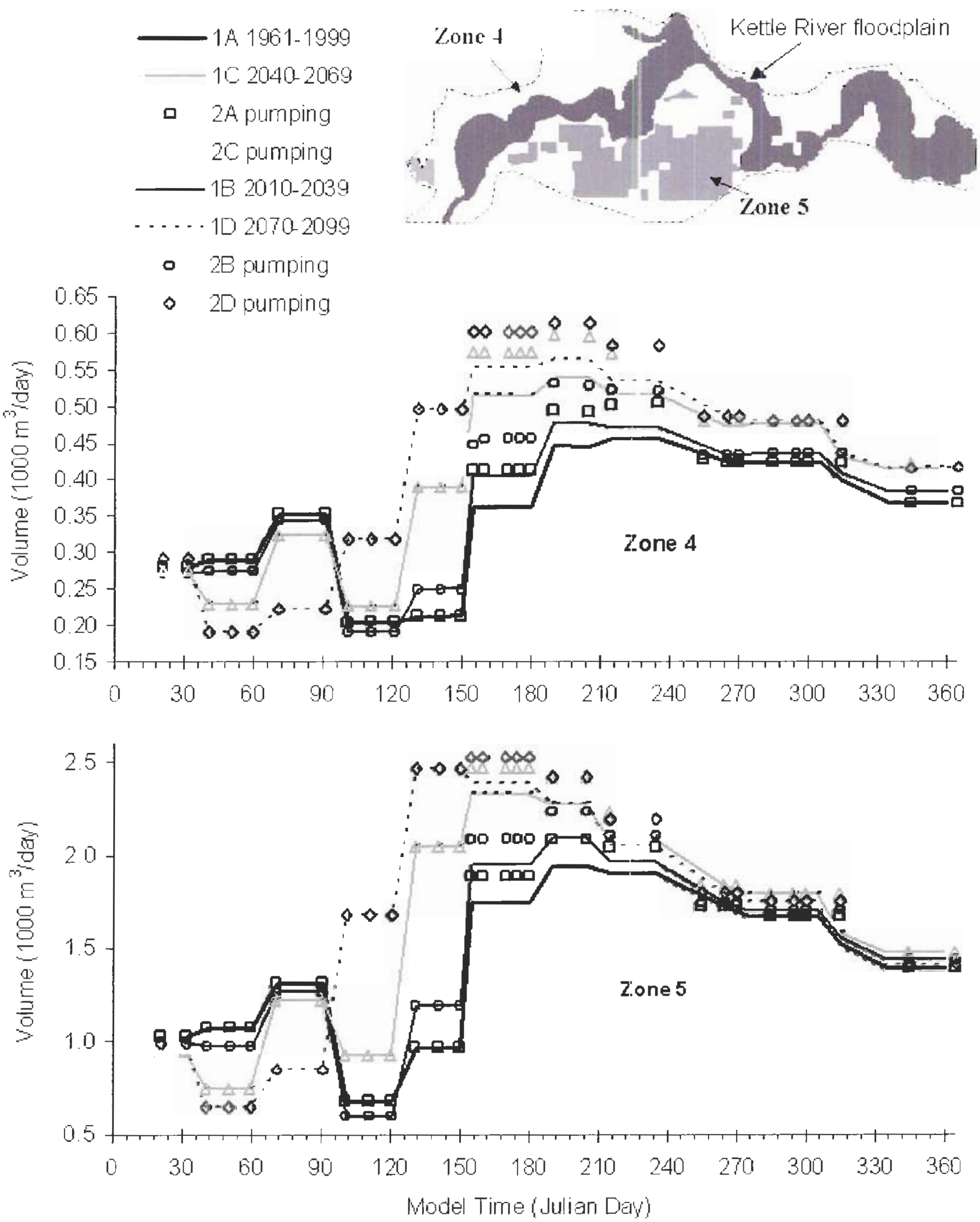

Figure (2) - 16: Recharge for Zones 4 and 5 comparing non-pumping to pumping and all climate scenarios. Symbol legend applies to both graphs. Note the different vertical scale on graphs.

During spring freshet on the Kettle River, the rise in river stage causes inflow of water to various ZBUD zones (after passing through the floodplain area). This excess water is stored in the aquifer. Mass balance calculations indicate that storage rates are less than 50% of inter-zonal groundwater flux, and 15 to 20% of river-aquifer flux. As river stage drops, the hydraulic gradient is reversed; water is released from storage and enters the floodplain zone where it eventually returns to river as baseflow. As most of the pumping water is lost to evapotranspiration on irrigated fields, there is a small reduction in the baseflow component to the Kettle River during the pumping period.

Currently, the peak of the hydrograph for the Kettle River occurs between day 100 and 150. In future climate scenarios the hydrograph peak is predicted to shift to an earlier date, although the peak flow will remain the same [Scibek et al., submitted]. The shifts in the river hydrograph are predicted to be much greater in 2040-2069 than 2010-2039, both compared to the modeled historical 1960-1999 time period.

Climate Change Impacts on Groundwater Levels

Within an annual cycle and between climate scenarios the results show different spatial and temporal distributions in groundwater conditions. Head difference maps were prepared to show only differences due to climate change between future climate scenario model outputs and present climate scenario model outputs (Fig. 17). Pumping effects were subtracted out because drawdown was identical in all climate scenarios (pumping rates were constant in all models for the pumping time period).

At present day, the flow patterns are influenced by river channel profile, and generally follow valley floor topography. In this aquifer, the effect of changing recharge on groundwater levels is very small compared to changes in timing of basin-scale snowmelt events in the Kettle River and subsequent shift in hydrograph [Scibek et al., submitted].

In the 2010-2039 scenario, water levels rise and fall with the river hydrograph at different times because of a shift in river hydrograph peak flow to an earlier date. The maximum aquifer water levels associated with the peak hydrograph are very similar to present climate because the peak discharge is not predicted to change, only the timing of it. Elevated water levels up to 30 to 40 cm persist along the channel and drain within a month. From late summer to the end of the year, water levels are similar to present conditions, with small increases observed due to the increase in recharge in areas away from the river channel. In the 2040-2069 climate scenario, the hydrograph shift is larger than in the 2010-2039 climate scenario, resulting in up to 50 cm change in groundwater levels.

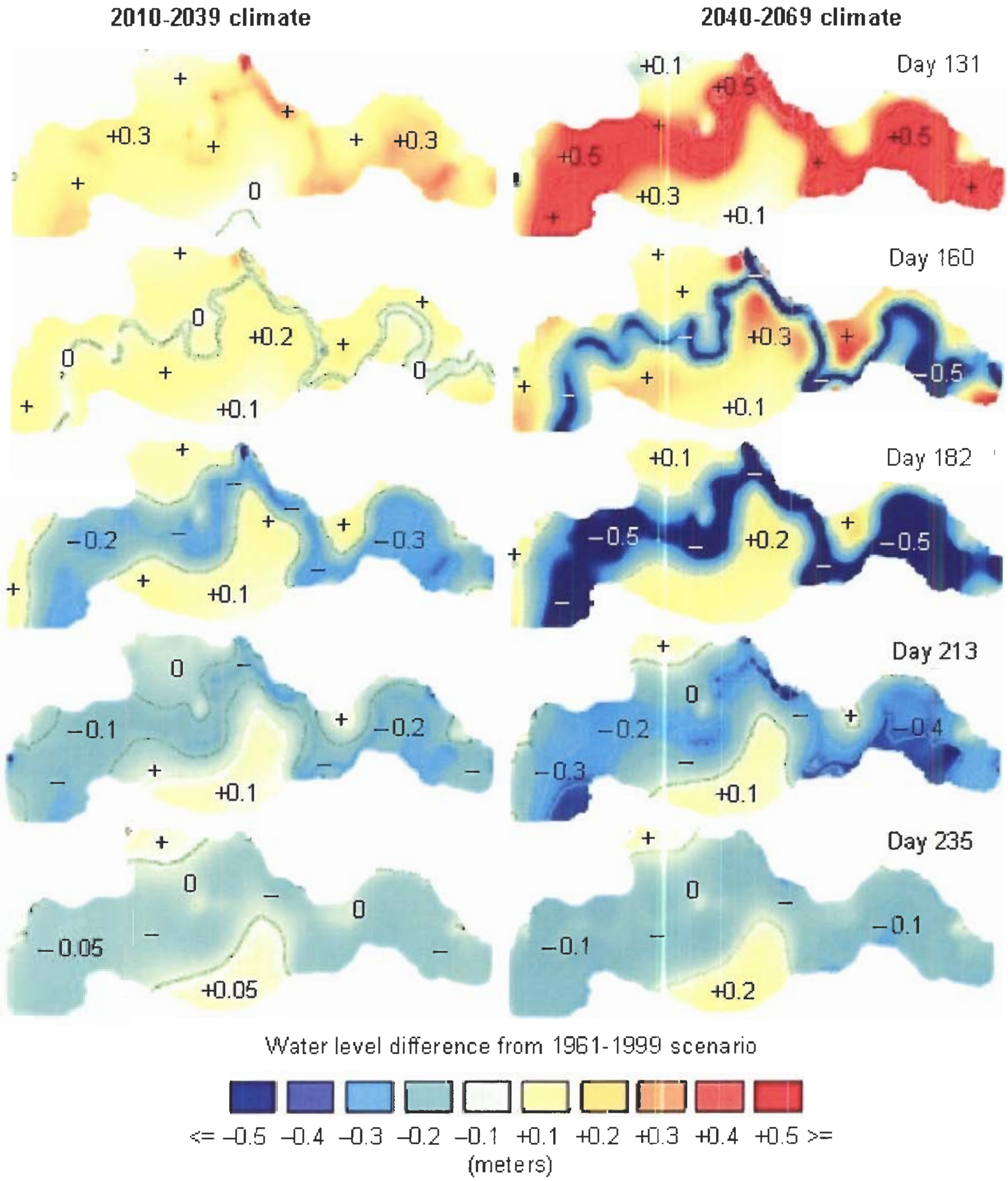

Figure (2) - 17: Water level differences (measured as head in layer 2 of the unconfined aquifer) between future (2010-2039) and present climate on left and between future (2040-2069) and present climate on right. Maps by time step in days 131 to 235. Contours shown are at a 0.1 m interval. The zero contour is dashed line. Darkest blue colours indicate values < -0.5 m (along rivers only). At day 101, difference map (not shown) has values within 0.1 m of zero.

Conclusions

In undertaking climate change impacts modeling, we rely on estimations of future climate as determined from imperfectly-calibrated downscaling models, which themselves are from similarly uncertain results of CGCM1 climate model. This inevitably leads to model uncertainty, as demonstrated in this study by the large calibration bias between the downscaled present climate and observed present climate, particularly during the summer months at Grand Forks where precipitation was underestimated by up to 40% compared to observed.

For precipitation predictions at this location, the choice of downscaling method was shown to be very important for interpreting predictions of GCM models, as GCMs do not directly model precipitation at a local site. There are many local controls on local precipitation (elevation, rain shadow effects, distance from ocean coast, etc.), which are not captured by a GCM. Two different downscaling methods were used and compared, the SDSM model and the K-nn model. Both are statistical models that link the climatic variables in CGCM1 climate model to observed precipitation and temperature at a specific location (the two local climatic variables). At the Grand Forks site, neither SDSM nor K-nn adequately models precipitation, and the two models differ in goodness of fit to observed precipitation in different months of a year. The SDSM model results were selected because of better fit in the spring season to observed precipitation, when much of the recharge to surficial aquifer is thought to occur. To overcome the limitations of the downscaling results, relative changes in climate parameters were used to adjust predicted historical climate data within a stochastic weather generator.

The HELP hydrologic model used in this study was found to be sensitive to depth of vadose zone (percolation layer), soil type, and saturated hydraulic conductivity (Ksat) of the vadose zone. Therefore, in order to achieve accurate results for recharge, it is important to capture the spatial variability of these key variables. Results indicate that Grand Forks receives between 10% and 80% of recharge from precipitation, depending on location within the valley.

The spatial distribution of recharge has consistently greater control on flow rates than temporal distribution of recharge. In this particular aquifer, the model is sensitive to recharge only away from river floodplain, and the maximum change in water table elevation is between 10 and 50 cm, but typically about 20 cm. Areas of the aquifer where temporal variation in recharge does not significantly affect model output are along river floodplains. There, water levels are almost entirely controlled by river water levels.

The predicted future climate for the Grand Forks area from the downscaled CGCM1 model will result in more recharge to the unconfined aquifer from spring to the summer season. That said, the river water level perturbation is much more important here, according to model simulations and sensitivity runs, than recharge perturbation over the aquifer area, because of the vastly different flow rates and volumes involved (river versus precipitation recharge), but nonetheless, the overall methodology used in the methodology will allow for similar studies to be undertaken on aquifers elsewhere.

3 GROUNDWATER-SURFACE WATER INTERACTION

Groundwater-Surface Water Interaction Under Scenarios of Climate Change Using a High-Resolution Transient Model

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Introduction

With increasing concerns surrounding global climate change, there has been growing interest in the potential impacts to aquifers. It is expected that predicted global changes in temperature and precipitation will alter regional climates and hydrologic systems. One of the expected consequences will be changes in recharge to regional groundwater aquifers, causing shifts in groundwater levels [Changnon et al., 1988;

Zektser and Loaciga, 1993]. Most research to-date has been directed at forecasting the potential impacts on surface water hydrology, while for groundwater hydrology, large regional and coarse-resolution models have been used to determine the general sensitivity of groundwater systems to changes in critical input parameters, such as precipitation and runoff [York et al., 2002, Yusoff et al., 2002], with only a few detailed investigations of small aquifers [e.g., Malcolm and Soulsby, 2000]. Of particular interest are coupled hydrologic systems, where changes in surface flow regime, and changes in recharge to groundwater interact to affect both groundwater and surface water levels.

The purpose of this study is to determine the potential impacts of climate change on groundwater levels in a small regional unconfined aquifer (34 km^2) by modelling groundwater flow in a transient three-dimensional numerical model, which is linked to river flow and has spatially- and temporally-distributed aquifer recharge. Scibek and Allen [submitted a] describe the recharge modelling methodology and impacts of climate change on recharge and groundwater levels, respectively. The objective of this work is to assess the groundwater-surface water interactions in the Grand forks aquifer and to assess the impacts of future climate scenarios.

Aquifer Model Development

The unconfined Grand Forks aquifer, located in south-central British Columbia (BC), Canada (Fig. 1), is contained within the mountainous valley of the Kettle River in BC along the Washington State (WA), United States border. At Grand Forks, the climate is semi-arid and most precipitation occurs in the summer months during convective activity. In the winter, much of the precipitation at high elevation is as snow, although the observing sites at valley bottoms record less snowfall. Groundwater is used extensively for irrigation and domestic use [Wei et al., 1994].

Within the Grand Forks valley, the Kettle River is a meandering gravel-bed river incised into glacial outwash sediments, and previous studies [Allen et al., 2003] demonstrated that the aquifer water levels are highly sensitive to water levels in the Kettle River. We suspect that any climate change in this region may cause shifts of the timing and amplitude of annual discharge hydrograph on the Kettle River, as discussed in a later section, consideration of impacts of climate change must necessarily consider both surface water and groundwater.

To construct a groundwater flow model, first, the valley shape was modelled using profile extrapolation, constrained by well lithology logs, and geostatistical interpolation. The valley was found to attain a maximum depth of approximately 250 m below ground surface, but typical sediment thickness is about 100 m. The stratigraphic sequences in the Grand Forks valley are poorly understood, particularly at depth. Approximately 150 well lithology logs are available for mostly shallow groundwater wells, and have been previously interpreted within the uniformly layered paradigm of hydrostratigraphy [Wei et al., 1994]. In other valleys in southern BC, the basal units are commonly silt, clay and gravel, overlain by thick glaciolacustrine silts [Fulton and Smith, 1978; Clague, 1981; Ryder et al., 1991], and capped by Holocene sandy and gravelly outwash and floodplain deposits and paraglacial alluvial fans.

The hydrostratigraphy was interpreted from selected high-quality logs of well boreholes. These have been used previously in groundwater well capture zone delineation and the logs have been interpreted by trained hydrogeologists. The interpreted layers were constrained by the Quaternary depositional history of the valley sediments. Hydrostratigraphic units were modelled in three-dimensions from standardized, reclassified, and interpreted well borehole lithology logs. Solid models were constructed using GMS software (v. 4.0) [Brigham Young University, 2002], converted to a five layer system underlain by solid bedrock, and imported into MODFLOW, as is typically done with complex multi-layer aquifer systems [Herzog et al., 2003]. Details of model construction are described in Allen et al. [2004]. Representative homogeneous and isotropic hydraulic properties were initially assigned to each layer, based on values determined from pump test data, but were later modified during model calibration. The model was calibrated to replicate the observed variation in groundwater levels in the floodplain of the Kettle River at Grand Forks, as described in a later section.

Hydrology of Kettle and Granby Rivers

The Kettle River system drains approximately 9,800 km^2 within BC, where effectively most of this study area is located (Fig. 1). The river crosses the US border at Ferry, WA, and loops back to the Canadian side at Carson, BC (Fig. 2). The valley widens near the City of Grand Forks, where the Granby River flows into the Kettle River. The Kettle River flows east through a narrow valley for about 10 km, turns south near Christina Lake, and crosses the US border at Laurier, and drains another large area before it flows into the Columbia River. The Granby River has a drainage area of 2,050 km^2 at its confluence with the Kettle River at Grand Forks BC.

In the Kettle River drainage area, the snowpack increases over the winter until early April, and melts between April and the end of June, with the end date of the snowmelt season varying from mid-May to mid-July. The hydrological response is extremely sensitive to seasonal variations in climate. During years with unusually warm winters the system shifts from a snowmelt-dominated regime to a regime where there is an increasing number of days of higher flows due to rain, but with a decreasing number of days of high flow due to snowmelt. The predicted warming trends in global, and also regional, climate, may impact the snowfall amounts and the duration of winter season, and may shift the hydrologic regime, potentially affecting hydrologically linked regional aquifers. In this study, we investigate these linkages of the Kettle River and the Grand Forks aquifer.

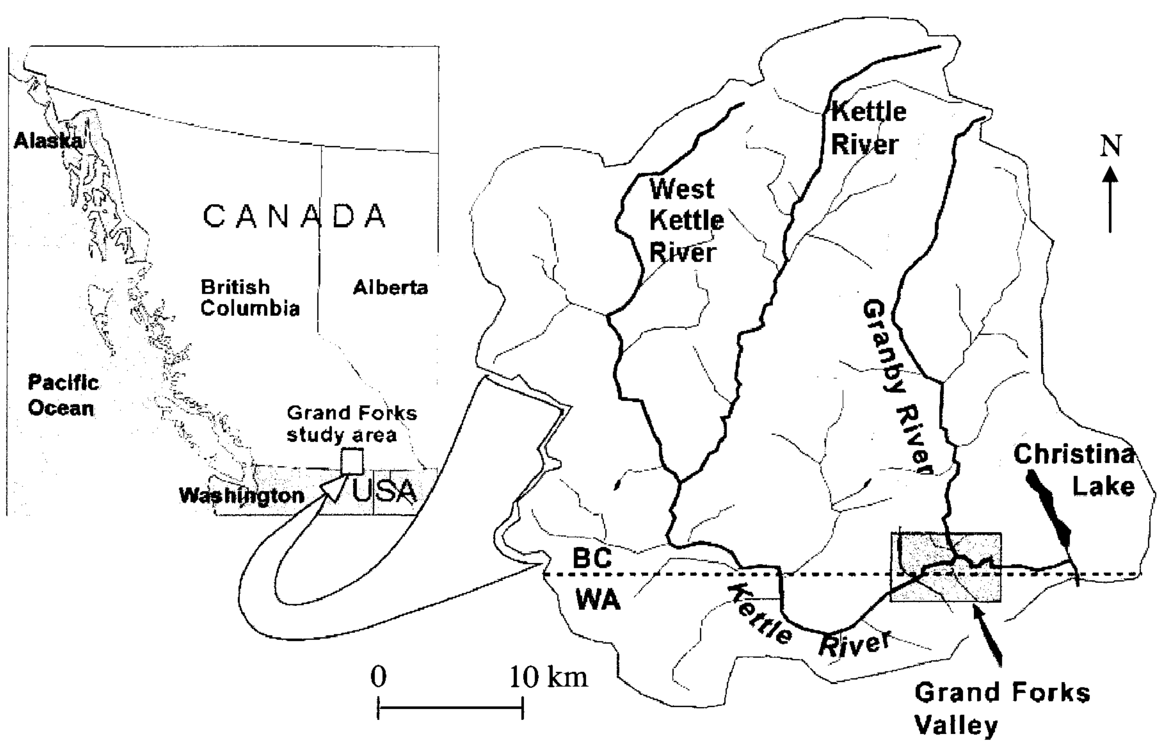

Figure (3) - 1: Map of the Kettle and Granby River drainage areas with inset maps show the study area in British Columbia, Canada.

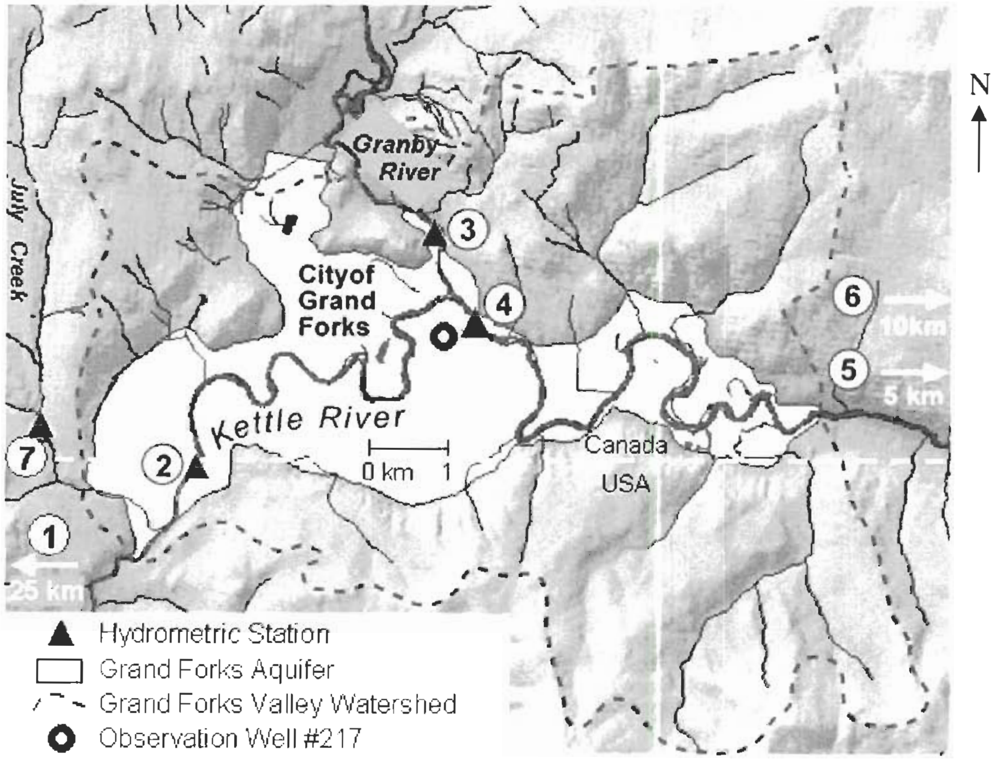

Figure (3) - 2: Grand Forks valley watershed and hydrometric stations near the Grand Forks aquifer. (1) Kettle River at Ferry, (2) Kettle River at Carson, (3) Granby River at Grand Forks, (4) Kettle River at Grand Forks, (5) Kettle River at Cascade, (6) Kettle River at Laurier, (7) Kettle River at Laurier.

Whitfield and Cannon [2000 a] analysed data from hydrometric stations in southern BC over two decades (1976-1985 and 1986-1995). The study determined these streams are currently snowmelt-dominated. Observed changes in Kettle River discharge between these two decades, is indicated on the polar plot shown in Figure 3, which shows a shift in peak flow to an earlier date, although the peak flow magnitude remains the same. Similar responses were observed in other streams in South-central BC (e.g., Similkameen River) [Whitfield and Cannon, 2000 a]. The low flow period now begins earlier in the summer and baseflow levels are lower in fall. In addition, flow is higher in the late fall due to rainfall during this period.

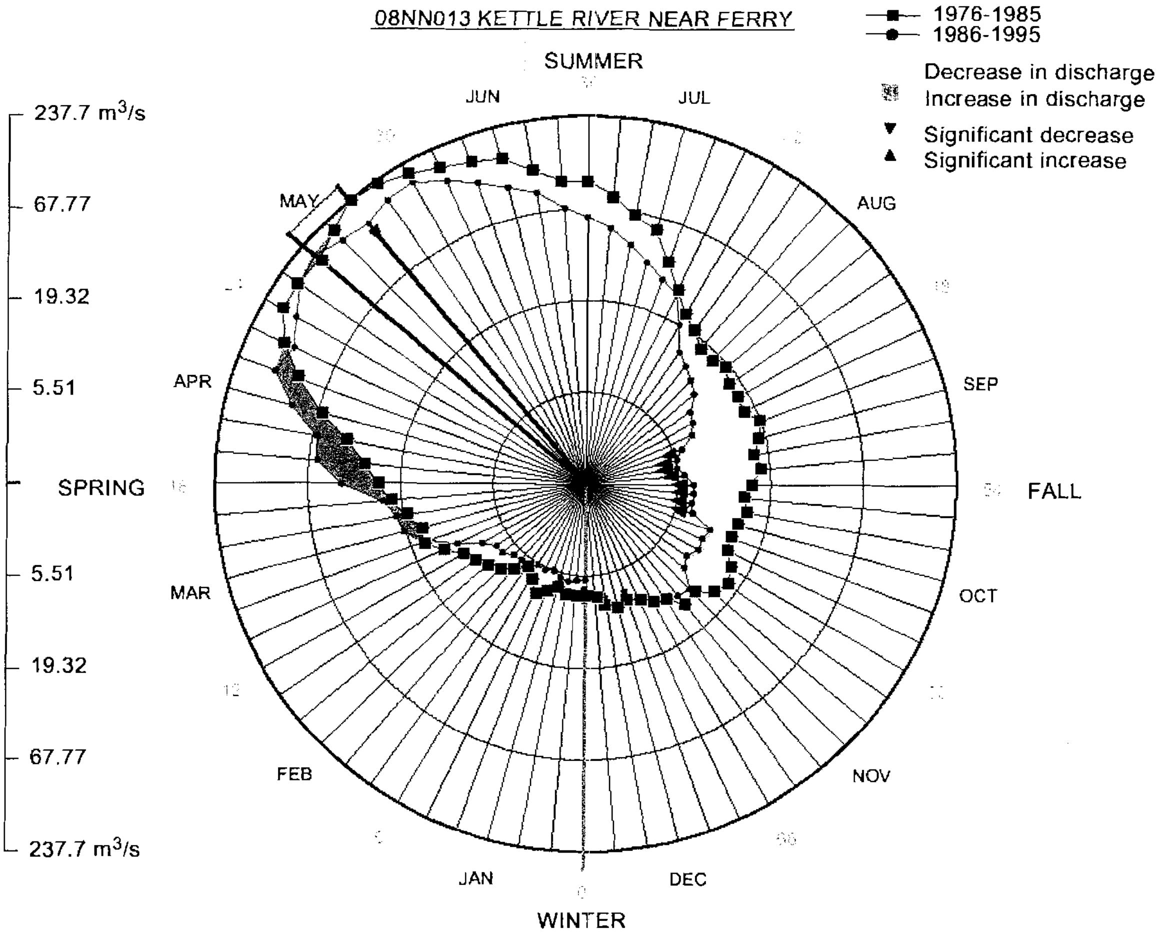

Figure (3) - 3: Observed changes in streamflow on Kettle River near Ferry, WA between 1976-85 and 1986-95. The shading between the two curves is dark when increased, and light when decreased. Up and Down arrowheads indicate statistically significant changes (0.05) as determined by the Mann-Whitney test [Leith and Whitfield, 1998]. The two radial lines extending beyond the circle indicate the highest flow in the first period (dark) and the second period (light).

River Discharge Rates in Grand Forks Valley

In order to model the interaction between groundwater and surface water in the valley, stage elevations are required as a function of time for each river node in the groundwater flow model for each climate scenario. The challenges in constructing the model were firstly, balancing the discharge volume in the valley, given that hydrometric stations are located outside the valley and have different periods of record; secondly, modelling basin scale discharge from downscaled Global Climate Model (GCM) outputs; and thirdly, accurately modelling stage variation in river branches such that stage could be linked to the groundwater flow model and used to predict impacts on groundwater levels.

Daily discharge records were supplied by Environment Canada and from the United States Geological Survey (USGS). As most river gauges record only water elevation, the discharge records are calculated from stage-discharge rating curves. Representative annual hydrographs, averaged for the period of record, were plotted for each hydrometric station. An example of the hydrograph at Ferry, WA, is shown in Figure 4. The available hydrometric stations in the valley have non-overlapping periods of record; the longest records are at Ferry (WA) and Laurier (WA) gauges on the Kettle River. Therefore, it is necessary to scale these discharge records to represent flow at points between these two gauges in the Grand Forks valley.

Using runoff as a hydrologic response allows for adjustments for differences in drainage areas. Runoff (R) is the volume of streamflow discharge (Q) over a period of time (t) divided by the drainage area (A_B) of the basin:

$$R = \frac{Q \cdot t}{A_B} \quad [1]$$

Discharge records from gauges along the Kettle and Granby Rivers were converted to 30-day runoff values. The Granby River basin has much larger runoff values than the Kettle River basin, suggesting greater precipitation in that region.

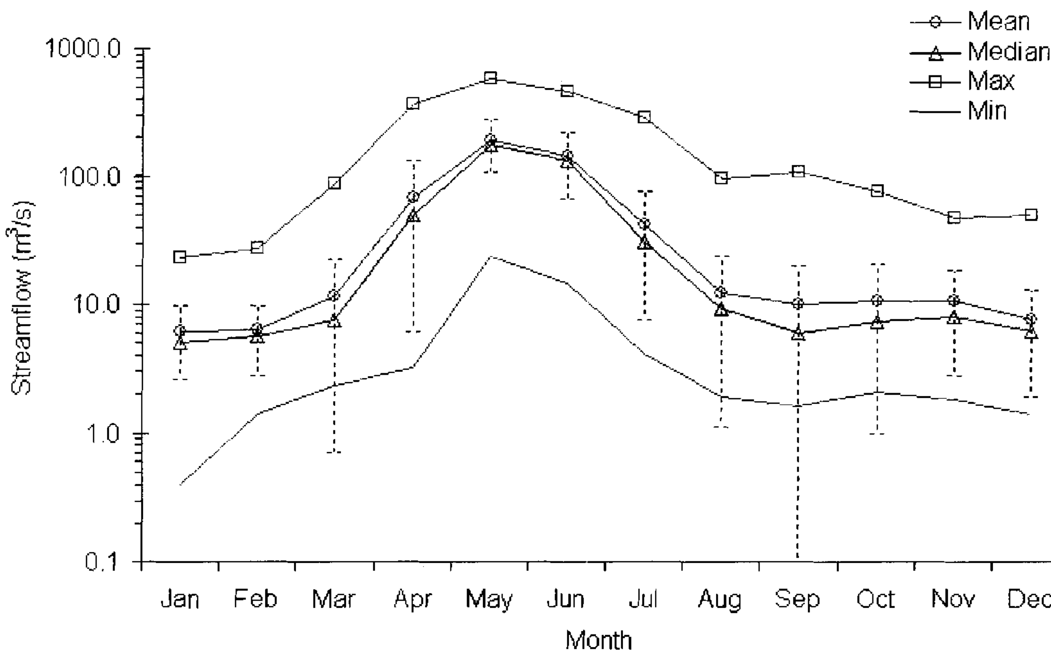

Figure (3) - 4: Monthly discharge statistics calculated from mean daily discharges for the complete period of record (1928–1996) for the hydrometric station on the Kettle River at Ferry, WA. Plotted are the mean, median, maximum, and minimum values. Also shown are standard deviations of monthly means (vertical bars).

To determine the runoff at a location downstream of a gauge, the observed daily flows at the upstream station were adjusted by the drainage area ratio of downstream/upstream stations, following methodology of Leith and Whitfield [2000]. The scaling factor was computed from the product of the ratio of basin areas at upstream station (A_1) and downstream station (A_2) locations along the river. Then, discharge at the downstream station (Q_2) was computed from discharge at the upstream station (Q_1) using the equation:

$$Q_2 = \frac{A_2}{A_1} Q_1 \qquad [2]$$

The streamflow records at Laurier were scaled (Equation 2) to represent the streamflow hydrographs in the Grand Forks valley downstream of the confluence of the Kettle and Granby Rivers. The upper section of Kettle River in the valley was then modelled using scaled discharge values from the gauge at Ferry (Table 1).

For the Grand Forks valley, the inflow discharge Q_{IN} can be separated into the flow components (Equation 3) of the two rivers, the tributary creeks, and baseflow from the aquifer in the valley (includes groundwater flow into the river channel, flow from drains and ditches and storm sewers). On an annual time scale, the inflow components for the Kettle River channel in the valley are:

$$Q_{IN} = Q_{RIVERS} + Q_{CREEKS} + Q_{BASEFLOW} + Q_{STORAGE(IN)} + \Delta Q = Q_{OUT} \qquad [3]$$

$$Q_{IN} \sim Q_{OUT} \qquad \text{where } \Delta Q \sim 0$$

assuming one year cycle has no long term trend (multi year). With no change in long term storage (ΔQ), each annual cycle completes the water balance so that inflow and outflow terms are equal on annual time scale. The outflow terms for Kettle River are either outflow from valley or water pumped out for water supply and irrigation, which is taken from the surficial aquifer:

$$Q_{OUT} = Q_{RIVERS(OUT)} + Q_{STORAGE(OUT)} \qquad [4]$$

Note that in Equation 4 on short time scales there is a storage term for the surficial aquifer ($Q_{STORAGE}$). In the aquifer, $Q_{Rivers} >> Q_{Creeks} >> Q_{Baseflow}$, and on an annual time scale, $Q_{IN} = Q_{OUT}$ for the valley.

The inflow from creeks in the Grand Forks valley watershed (95 km^2) to the rivers must be estimated using a scaling approach (Equation 2). Here, the mean annual discharge in the July Creek (Figure 2) catchment (45 km^2) is known from records, and is representative of local climate and hydrology. Thus, its discharge is increased by a factor of 2.09 (ratio of 95/45) to arrive at estimates of minimum, maximum and mean discharge in the Grand Forks watershed as shown in Table 2.

Table (3) - 1: Selected Kettle River and Granby River hydrometric stations.

Station	Reference Number	Basin Area (km^2)	Conversion	Scaling Ratio
Kettle River at Ferry	USGS 08NN013	5750		
Kettle River at Carson	WSC 08NN005	6730	Ferry → Carson	1.1704
Granby R. at Grand Forks	WSC 08NN002	2050		
Kettle River at Grand Forks, below confluence with Granby	estimated	6825 + 2050 = 8875	Kettle River + Granby River	
Kettle River at Cascade	WSC 08NN006	8960		
Kettle River at Laurier	USGS 08NN012	9840		

Table (3) - 2: Estimated contribution of discharge from drainages within the Grand Forks valley watershed, scaling up from the July Creek catchment.

Annual Discharge (m/s)	July Creek (m^3/s)	Grand Forks Watershed (m^3/s)
Minimum	0.00686 x 2	0.0137
Maximum	2.06 x 2	4.12
Mean	0.32 x 2	0.64

The mean annual discharge of the Granby River is 30.5 m^3/s, and for the Kettle River, upstream of Grand Forks, it is 44.3 m^3/s, both adding to 70.8 m^3/s. Downstream from this confluence, mean annual discharge is 72.8 m^3/s as measured at Cascade hydrometric station. This small discrepancy comes from different periods of record available at those three locations. Therefore, at the confluence of these rivers, the Granby contributes approximately 40% of the flow, and the Kettle contributes 60% of the flow to the Kettle River. The ratio of discharge of 0.69 from the Granby to the Kettle River varies from year to year. In most years, at low flow in August, the Kettle River maintains a discharge of between 10 and 14 m^3/s, compared to a minimum discharge of 0.0137 m^3/s from creeks in Grand Forks watershed and baseflow from the aquifer. Thus, during the mean flow or high flow conditions, the small tributaries contribute only 0.64 to 4.12 m^3/s daily discharge to the larger Kettle River, within the extent of the Grand Forks aquifer, or approximately 1% of the combined Kettle and Granby River discharge.

Simulating River Flows of the Kettle and Granby Rivers

The BRANCH model is a one-dimensional flow model developed and validated by the USGS [Schaffranek et al., 1981]. The model is intended for broad operational use to compute unsteady flow and water-surface elevation (stage) of either singular or interconnected channels. The time-dependent variables are the flow rate and the water-surface elevation. Water-surface elevations and flow discharges are computed at segment nodes and branch junctions.

A new user interface was developed for the BRANCH code, where all inputs and outputs are included in a single spreadsheet file. Model parameters and inputs were read either directly from spreadsheets, or optionally from old BRANCH format text files. A new module was written to allow for hydrograph generation to create boundary value data series in any time increments to simulate the hydrograph wave form based on monthly values. Mapping of the channel network into a raster grid as defined by groundwater flow model implemented in Visual MODFLOW v 3.1.84 [Waterloo Hydrogeologic Inc., 2004]. This divides the channel into segments, and uses BRANCH output to update the MODFLOW boundary value file for specified-head boundary schedules for any number of cells. The new version of BRANCH was verified with USGS sample data.

The model was applied to 26 km length of the Kettle River channel, including a small section of the Granby River (about 1 km). The channel sections of the Kettle and Granby Rivers with BRANCH schematization are presented in Figure 5a. Boundary conditions were specified at three external nodes, and river stage was computed at 67 channel cross-sections [British Columbia Environment, 1992] as shown in Figure 5b. It is important to note that although the cross-section spacing along the Kettle River is dense (150 to 600 m with an average approximately 400 m), the river channel geometry varies greatly with location. There is also a lack of consistency in high-water mark surveying along the cross-sections. Thus, due to lack of more information, it was assumed during calibration that not all the high-water level scour or debris on channel banks was caused by the same high flow at all points along the channel. Limitations of the model also include a lack of accounting for channel storage, variations of channel roughness with stage, or backing up of water along un-surveyed sections of the channel, which could impact the surveyed locations. Therefore, neither the surveyed high-water marks nor the modelled stages are without error. At low flow, there are small rapids in various places along the river channel, violating the assumptions of the model in some small sections, but this should not matter for most of the river sections as there are no rapids there.

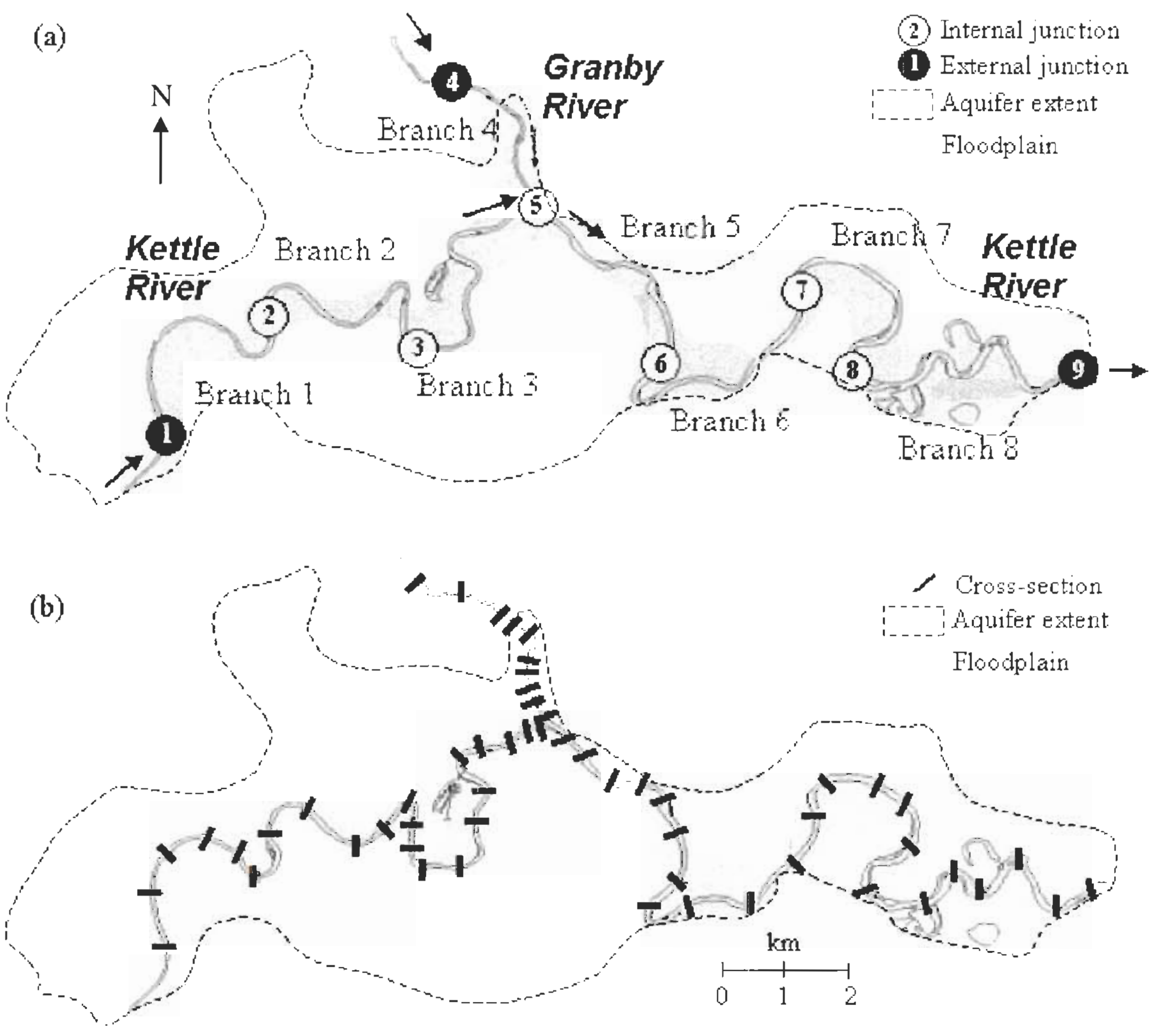

Figure (3) - 5: (a) River branches in the BRANCH model, (b) location of Kettle River cross-sections 1 to 67. In the upper panel, the numbers in the dark circles indicate external junctions and the numbers in the light circles indicate internal junctions. The dark shading shows the extent of the river floodplain, and the dashed line the extent of the aquifer.

Stage-Discharge Curves

Notwithstanding the limitations of BRANCH, stage and discharge (rating curves) were calculated for all river cross-sections at 1-minute time intervals over the specified number of 10000 time steps. The input consisted of a rising river discharge hydrograph from baseflow to near peak flow, similar to that observed in early freshet, for a typical range of discharge values. Tabular and graphical output was specified at coarser time steps of 60 to 120 minutes, but finer output steps were used for calibrating initial stages at early time steps of model runs. Stage-discharge plots were created from scatterplots of computed stage and discharge for each cross-section. Rating curves were fitted with a simple power law function ($y = ax^b+c$). The shapes of these curves are compared to historical rating curves from hydrometric stations in the Grand Forks Valley (Fig. 6a).

High water marks were added to stage-discharge plots as horizontal lines and the fitted curves were extrapolated to intersect the high-water mark lines at typical flood level discharges (Fig. 6b). If the modelled rating curves deviated greatly from the high-water mark and the cross-section geometry indicated that modelled stages were not reasonable when compared to other nearby cross-sections with similar geometries, the rating curve equation was adjusted and the curve fitted to intersect the high-water mark. As described in the following sections, river stage hydrographs at all cross-sections were imported to the groundwater model as nodal boundary conditions by linearly interpolating between cross-sections.

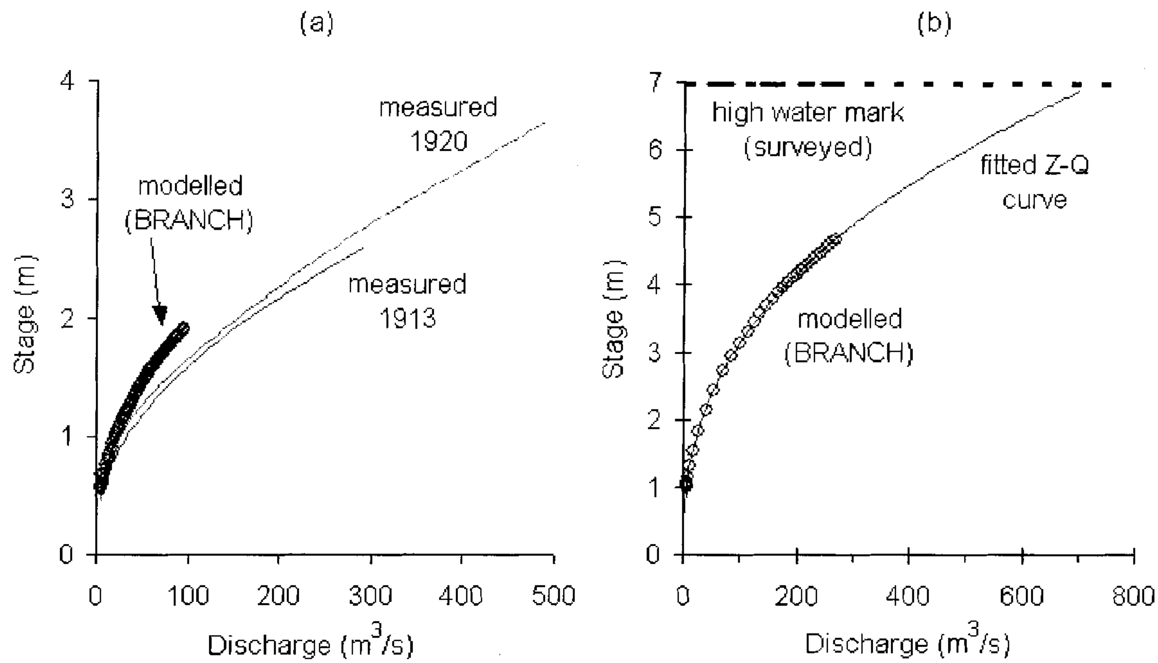

Figure (3) - 6: (a) Stage-discharge curves derived from historical records (measured in 1913 and 1920, from Water Survey Canada records [Environment Canada, 2002]) and modeled by BRANCH for gauges on the Kettle River in Grand Forks valley near US-Canada border, (b) BRANCH output, surveyed high water mark, and fitted rating curve for Kettle River channel in Grand Forks near confluence with Granby River.

Type of River Boundary Condition in MODFLOW

The correct selection of boundary conditions is a critical step in model design [Anderson and Woessner, 1992]. Transient simulations are needed to analyse time-dependent problems, such as the impact of climate-change induced shift in river hydrographs on the water levels in the Grand Forks aquifer. Boundary conditions influence transient solutions when the effects of the transient stress reach the boundary, and the boundaries must be selected to produce a realistic simulated effect. The MODFLOW model contains two packages that account for leakage to and from rivers. The River package allows rivers to be represented with a stage fixed during a stress period, with leakage to and from the aquifer [McDonald and Harbaugh, 1988]. It requires an input value for streambed conductance to account for the length and width of river channel, the thickness of riverbed sediments, and their vertical hydraulic conductivity. New versions of Visual MODFLOW include a Streamflow-Routing Package, which allows leakage to and from the stream, but it also maintains mass balance between the river and the aquifer. The Streamflow package assumes a very simplified uniform rectangular geometry for the river channel, which simplifies greatly the non-linear stage-discharge relation.

As was determined from the water balance in the valley, the river discharge in both the Granby and Kettle Rivers will not be measurably affected by inflows from small catchments in the Grand Forks Valley. Similarly, based on previous steady-state groundwater flow modelling [Allen et al., 2003] rivers are not measurably affected by baseflow from the aquifer. Thus, the combined aquifer and tributary contribution to the rivers have very small effect on Kettle and Granby River water levels. In contrast, the river water levels have a strong effect on groundwater levels in the aquifer. Therefore, the rivers can be represented as specified head boundaries, and the head schedules will represent the modelled river stage in transient Grand Forks aquifer model.

The bottom sediments of the Kettle and Granby Rivers above the Grand Forks aquifer consist of mostly gravels, with very few fine sediments. In effect, the aquifer is in direct contact with the river channel and there is no impediment to flow. The constant head nodes do not require any conductance coefficients, and thus, we assume perfect hydraulic connection between the river and the aquifer. The river can leak and receive water to and from the aquifer, but the river stage will not change as a result of such interaction. In other words, the river will act as an inexhaustible supply of water and will influence the aquifer water levels, but the aquifer will not have any effect on river discharge and stage. The head is held at a constant value for the duration of a time step, but changes to a different value with successive times.

Using BRANCH Output as Boundary Conditions in MODFLOW

The calculated rating curves, together with automated mapping of river water elevations to the groundwater flow model, allows for modelling seasonal variation of groundwater levels and their sensitivity to changed river hydrographs. Modelled discharge hydrographs were converted to river stage hydrographs at each of the 123 river segments, each roughly 200 to 250 m in length, and interpolated between known river channel cross-sections that have stage-discharge curves. River channels were represented at high grid density (14 to 25 m) in MODFLOW model. River segments were mapped onto MODFLOW cells in a GIS system (to mid points of cells), providing a database link between river water levels and appropriate river boundary cells. For each segment, the program located the nearest upstream and downstream cross-section location, and the stage-discharge rating curve for that cross-section was used to calculate water elevation from discharge. River water elevation was interpolated between cross-sections with fitted channel profile. The program then updated the appropriate boundary file of Visual MODFLOW. River stage schedules along the 26 km long meandering channel were imported at varying, temporal resolution (1 to 5 days) for every cell location independently. The channel width of Kettle River was 2 to 4 cell lengths at most locations. The actual thalweg, or water-filled and flowing channel width, may be less than two cell widths during low-flow months, but this schematization does not adversely affect the groundwater flow model.

Adjusting the River Elevation Profile

The DEM (20 m grid) was rather inaccurate in the valley and river floodplain elevations were too low in many places and the river channels poorly defined. The channel bottom elevation profile, representing the minimum elevation at each cross-section along the length of the river (Fig. 7), has a jagged appearance because there are local depressions in the river channel, or perhaps due to surveying errors. It would be expected that channel bottom would decrease or remain level in a downstream direction. This inconsistency of minimum channel bottom elevation profile caused problems in MODFLOW because the ground surface digital elevation model (DEM) and the river channels did not correspond to the surveyed channel bottom elevations. Thus, these had to be modified along river channels. River water elevation was calculated by adding river stage, computed from stage-discharge curve, to the channel bottom elevation. Consequently, the river channel bottom profile was smoothed out to ensure that calculated minimum and maximum stage were always decreasing downstream.

River stage should also be below local floodplain elevation, where “floodplain” is the area mapped along the river channel (on floodplain maps of the Kettle River) that would be flooded in a 20 year flood (or similar measure), because we did not attempt to model such extreme events in our model. MODFLOW layers were edited along all river channels to put all constant head boundary cells in first layer (gravel) of the model. The channels were also deeper than on the original DEM surface of the valley, but were similar to the surveyed channel profiles.

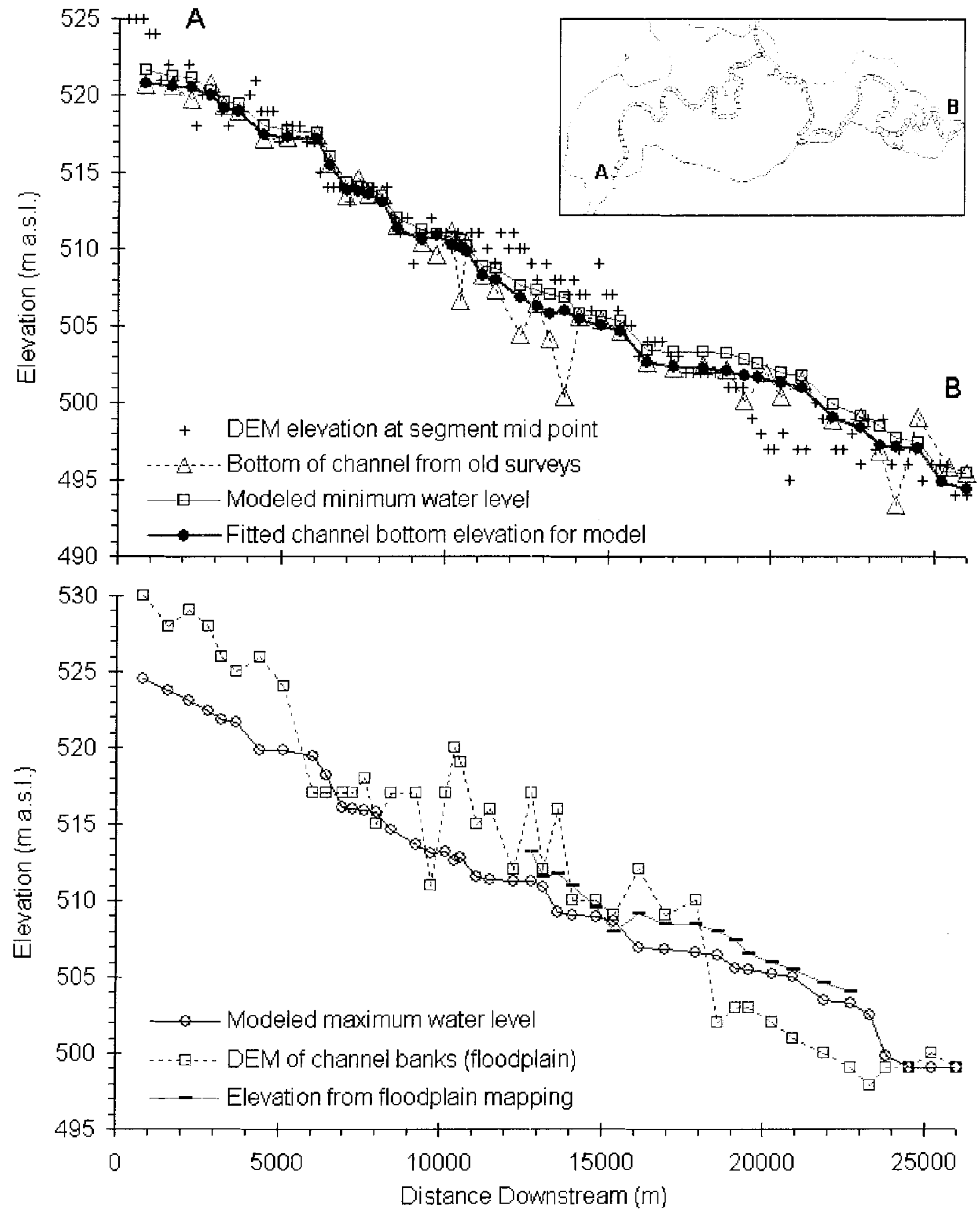

Figure (3) - 7: Elevation profile of the Kettle River. Top graph shows digital elevation model (DEM) at segment mid point, surveyed channel bottom elevations, modeled minimum water level flood, and fitted channel bottom. Bottom graph shows DEM of channel banks, modeled maximum water level and elevation from floodplain mapping.

Downscaling of River Discharge (Historical and Predicted)

Models for streamflow generation from watersheds can be calibrated to present conditions, and extrapolated to predict future conditions. These include physically-based watershed models, empirical or statistical models relating hydroclimatic variables to streamflow, and empirical downscaling models, where local or regional-scale variables (e.g., streamflow), which are poorly described by coarse-resolution GCMs, are related to synoptic- or global-scale atmospheric fields [Landman et al., 2001]. Beersma et al. [2000] showed climate scenarios useful for hydrologic impacts assessment studies. Climate downscaling techniques are treated in more detail by Hewitson and Crane [1996]. A review of applications of downscaling from GCM to hydrologic modelling can be found in Xu [1999].

The National Center for Environmental Prediction (NCEP) maintains a Reanalysis Project database [Kalnay et al., 1996], which provides large-scale climate variables that can be used to define analogs with GCMs for climate modelling purposes. Data from the NCEP/NCAR Reanalysis Project were extracted and used as historical analogs to make the climate-hydrology linkage using a technique suggested by Zorita and van Storch [1999]. Climate model output from the Canadian Global Coupled Model (CGCM1) [Flato et al., 2000] for the IPCC IS92a greenhouse gas plus aerosol (GHG+A) transient simulation were used to project results into the future. The meteorological data included 7-day sliding average of sea-level pressure, 500-hPa geopotential height, and 850-hPa specific humidity, 1-month sliding average of 500-hPa geopotential height and 850-hPa specific humidity, and 4-month sliding average of 850-hPa specific humidity. Three scenarios were generated using the calibrated downscaled model: current climate

(1960-1999), 2020's climate (2010-2039), and 2050's climate (2040-2069). The current climate statistics are based on 40 years of record, while the future climate scenarios represent three steps, each step representing an average of a 30 year period.

The dimension of the large-scale climate dataset was reduced using principal component analysis (PCA). A k-nearest neighbour analog model [Zorita and von Storch, 1999] was used to link principal component scores (explained variance > 90%) of the climate fields with the maximum temperature, minimum temperature, and precipitation series (of NCEP dataset). The PCA linked the climate fields over BC and the eastern Pacific Ocean with daily discharge values for Kettle and Granby Rivers. The analog modelling approach has the advantage of simplicity and comparable results to other more complex models. It also offers a simple method for controlling model fit and the time structure of the simulated series. The end product is sets of daily discharge data at the three sites for the simulated 1962-2100 period: Kettle River at Ferry (WA), Granby River at Grand Forks (BC), and Kettle River at Laurier (WA). The discharge data set for present climate scenario was truncated to 1971-2000 period (30 years) to make it the same length as the modelled river discharge for future climate scenarios. In the groundwater flow model, the base case (present climate) river hydrograph is the mean hydrograph for the 1971-2000 period (the original dataset 1962-2000 was also downscaled from 1962-2000 climate model runs from the GCM), while the downscaled climate for generating recharge to the aquifer in the groundwater flow model is based on GCM climate scenario output for the 1960-2000 period.

Output from the model consists of GCM downscaled discharge values for the three analog models with 7, 14, and 24 nearest k neighbours. There are two sets of results that correspond to different scaling factors being applied to the data: 1) "variance inflated" data have been scaled so that the variance of the simulated discharge values for the 1971-2000 period matches the variance of the observed values (i.e., overall variability is preserved), 2) "mean inflated" data have been scaled so that the mean values match (i.e., volume is preserved). There is no unique ideal solution, and the choice of scaling method depends on particular application of climate model results. The mean inflated discharge predictions were applied to groundwater model river boundary conditions.

The GCM gives one possible realization of simulated climate given historic forcings. The poor fit between the downscaled and observed hydrograph for 1971-2000 (Fig. 8) can mostly be attributed to biases existing between the GCM simulated climate fields and the observed climate fields from the NCEP-NCAR Reanalysis. The downscaled CGCM1 data underestimate temperature in the late winter and early spring periods and overestimate temperature in the late fall and early winter periods. Consequently, the onset of freshet is delayed. Wilby and Wigley [1997] demonstrated that downscaled climate scenarios are sensitive to many factors, including the choice of predictor variables, downscaling domains, season definitions, mathematical transfer functions, calibration periods, elevation biases and others. Although the output may be adjusted, there is a tradeoff between discharge time series "smoothness" and accuracy of modelled peak flows. This was expected and may be inevitable given the state of GCMs at present. The model bias is similar for all three hydrometric stations, but the model bias is greater for median discharges than for mean discharges. Therefore, only mean hydrographs were considered in future analyses.

Where the model bias is unacceptable, the downscaled results could be used as a basis for adjusting the observed historical hydrograph to match the simulated changes. However, such an approach might be hard to justify, especially for the future scenarios, and the GCM bias should be explicitly shown, along with the resulting impact on the subsequent hydrologic simulations [Whitfield et al., 2002] – see Fig. 8. The comparisons of impacts of future climates is then always between the unadjusted GCM-driven hydrologic simulations for future time periods and those for the baseline period.

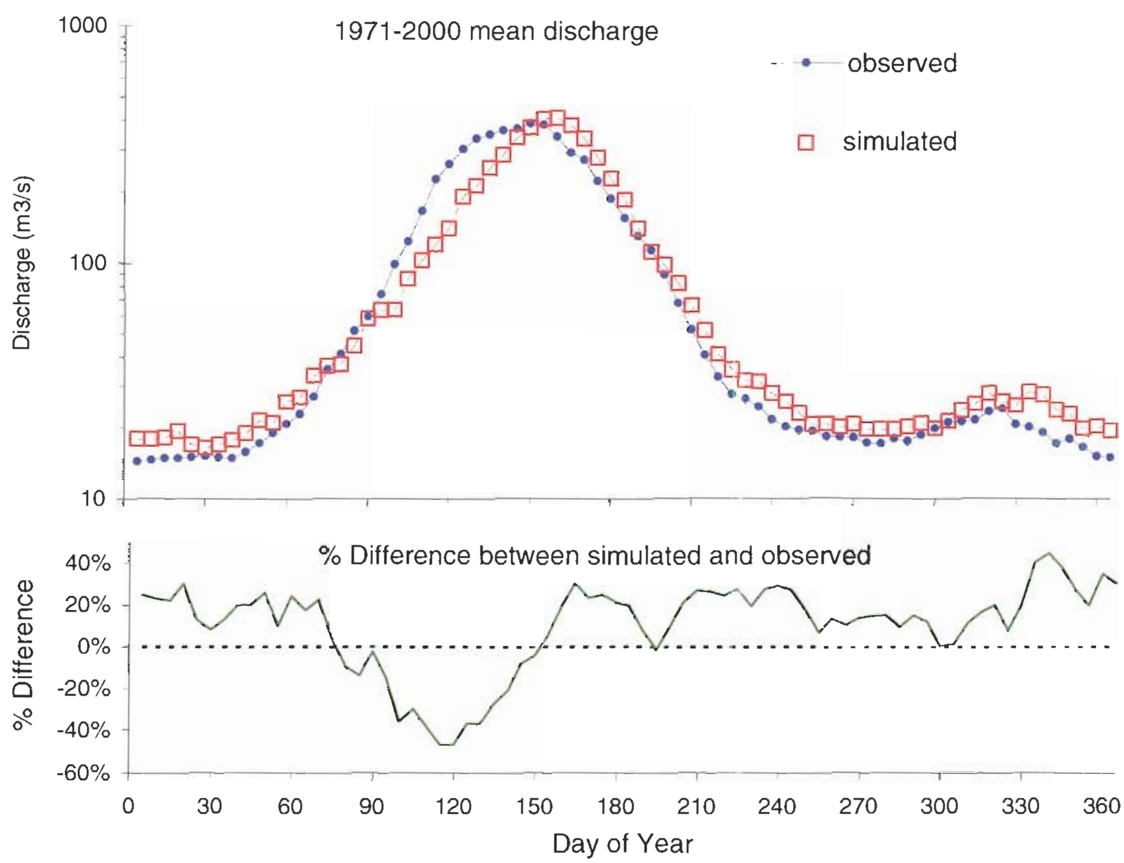

Figure (3) - 8: Observed and simulated discharge at Ferry (WA) on Kettle River, downscaled from CGCM1 showing model bias.

The hydrographs were generated for the climate scenarios, and graphed to compare the shifts in hydrograph shape (Fig. 9). Taking into consideration the model bias (modelled to observed, for present climate), the predicted changes in hydrograph are between the modelled present climate and the modelled future climate scenarios. In the future climate scenarios the hydrograph peak is shifted to an earlier date, although the peak flow remains the same. The recession part of the curve shifts rather evenly to an earlier date, but the rising part of the curve (day 60 to 120) has more chaotic change and overall less of a temporal shift. In the late summer and early autumn (days 250 to 320), the low flow period on the river shows a decreasing trend to lower discharges, from present to future climate scenarios. In other words, the models indicate that the minimum discharges on Kettle River will continue to decrease, which may be a problem for water supply in the future. There is also a significant increase in winter discharge in the future climate scenarios, most likely caused by an increase in rain and snowmelt volumes during the winter under warmer climate scenarios. Changes to the river hydrograph are predicted to be much larger for the 2040-2069 scenario than for the 2010-2039 scenario, compared to the modelled 1971-1999 period. The Kettle River and the Granby River had very similar responses to modelled climate change.

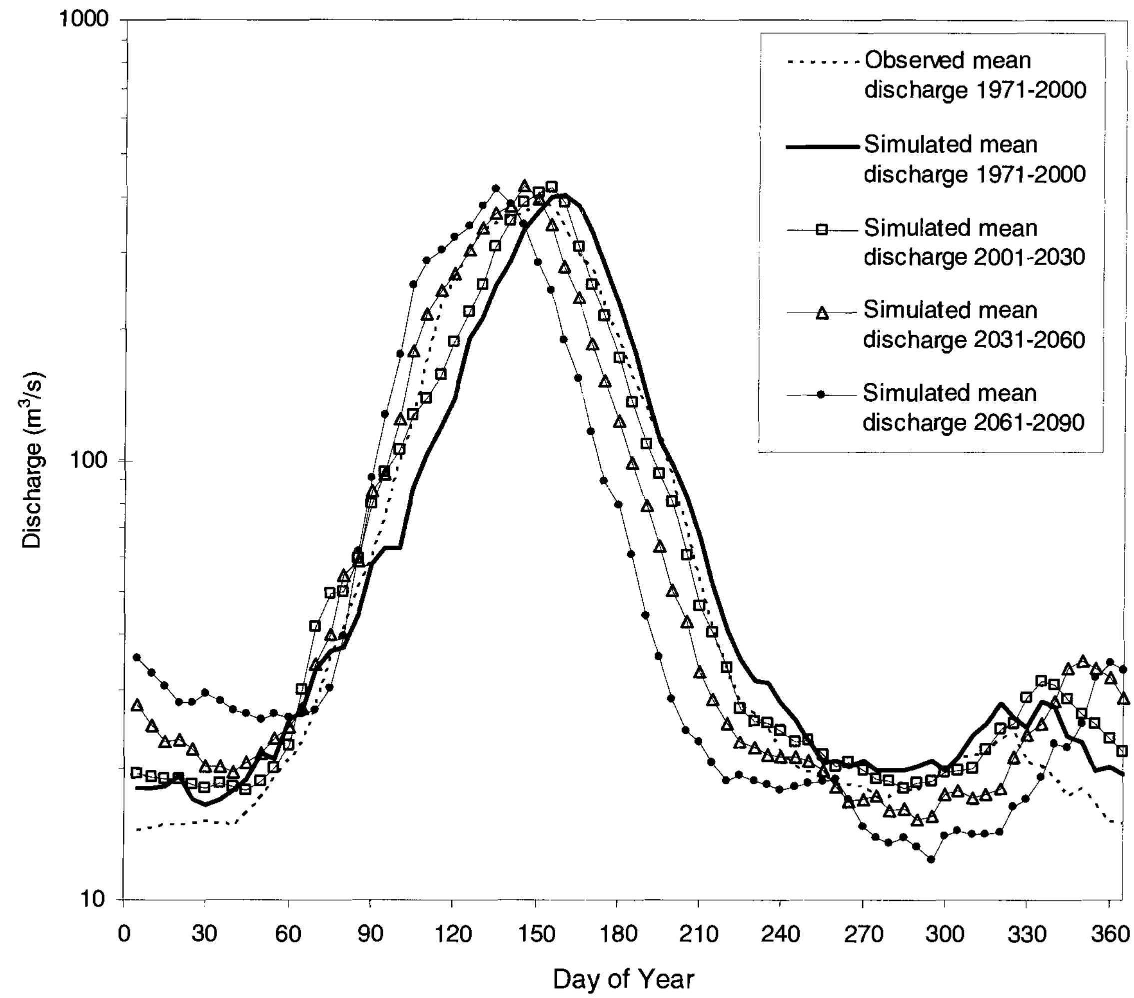

Figure (3) - 9: Predicted discharge in Kettle River at Laurier, WA, modelled using statistical downscaling model and comparing to observed discharge in last 30 years [Environment Canada, 2002].

Groundwater Model Calibration Results

The base case transient model simulated groundwater flow under present climatic conditions (1960-1999 scenario). Two models were created, one with pumping wells turned on and one without pumping of groundwater. Only large production wells within the aquifer were considered, and average mean daily pumping rates were assigned and activated during the summer months when groundwater wells supply irrigation demand in the valley. The calibration process was carried out on the non-pumping model, which is more representative of static groundwater levels.

The calibration graph for Observation Well 217, one of the BC provincial monitoring wells in Grand Forks shown in Figure 10, displays the observed long term monthly mean water elevation and modelled groundwater elevation after model calibration (present climate scenario). The groundwater levels in the observation well were taken on the last day of each month, then averaged for the period of record (1974-1996), and graphed on last day of each month. Also shown are observed and simulated discharge hydrographs for the nearby Kettle River at Carson for the corresponding time period. There is a regular seasonal pattern to groundwater levels, similar to the stage hydrograph of the Kettle River. The groundwater level in the well varied between 1 and 1.8 m over the period of record, whereas the river experienced stage fluctuation of 3 meters (Fig. 10). The mean monthly water table elevation varied only by about 1.0 ± 0.2 m. The shape of the well hydrograph is similar to the Kettle River hydrograph, but the amplitudes of seasonal fluctuations are damped, which would be expected at wells some distance away from the river channel.

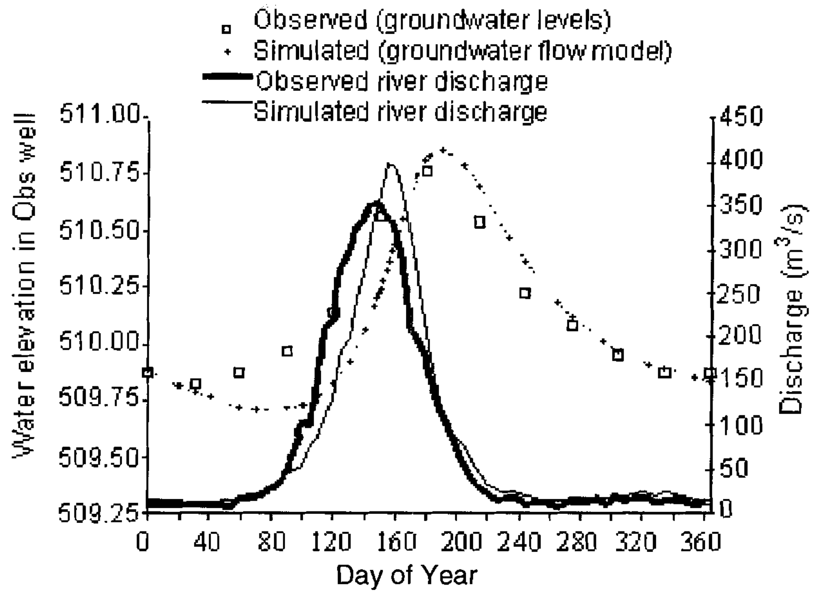

Figure (3) - 10: Mean hydrograph of water table elevation (total head) in Observation Well 217 in the Grand Forks aquifer for period of record 1974-1996, and estimated water surface elevation of Kettle River at Grand Forks (1971-2000) approximately 400 m from well 217 and Kettle River discharge.

The vertical offset of the hydrograph was calibrated by adjusting bottom elevation of river channels (see Fig. 7) around the observation well location, and by allowing for 0.2 m error in the absolute elevation of the observation well (top of casing). The water levels are measured with 0.01 m accuracy by a water level recorder and datalogger. The amplitude of the hydrograph (peak) was calibrated by changing specific yield (Sy) and horizontal hydraulic conductivity (Kxy) values in layers 1 and 2 (surficial gravel and sand aquifer) of the model. As Sy changed from 0.04 to 0.20, the amplitude of the hydrograph decreased and the slope of the decreasing hydrograph also flattened, ending with higher groundwater levels at the end of the year. The same effect was obtained by lowering Kxy and keeping Sy constant. A phase shift of the hydrograph occurs due to the delay of groundwater flow across the distance from the Kettle River (in 3 directions). It was calibrated by allowing for river model bias (from downscaled CGCM1 linked to river discharge), and by changing Kxy and Sy.

The results of the transient groundwater flow model showed lower than observed groundwater elevations at Observation well 217 in the early spring season. It is important to recognize that the river discharge hydrographs do not show the stage of the river. The BRANCH model, which was used to compute stage-discharge elevations, assumes ice-free conditions in the river, and only ice-free river stages were used in stage-discharge curves in the past for this river. It is well known that in Canadian rivers that partially freeze in the winter. The additional friction of floating ice increases stage throughout winter, resulting in higher actual river stage than would be predicted by a given discharge. The modelled water levels do not account for this effect because the groundwater model uses modelled river stage, which is computed from river discharge, without accounting for ice effects. Therefore, it is expected that modelled groundwater levels would decline from day 1 to day 60 when river discharge begins to increase due to snowmelt. In reality, the observed river stage is probably higher in spring due to ice damming and icing within the channel, but still conveys the same discharge.

The groundwater model is calibrated to the modelled (not observed) discharge and stage in the Kettle River. If there was no bias in the river model (observed to modelled present), the modelled well hydrograph would match the observed hydrograph. The modelled peak of groundwater was maintained at a level slightly higher than observed to account for this positive bias in modelled versus observed river discharge (and thus stage). Similarly, the modelled hydrograph is shifted to a later date. The calibrated model is also shifted by the same number of days to account for this bias. The groundwater model is very well calibrated at the location of observation well 217. However, this does not mean that it is well calibrated for other regions of the aquifer. Calibration residuals for static water levels had an acceptable error distribution, but residuals tended to be high near the model boundaries, which might be anticipated due to lack of physical data in these areas with which to constrain the conceptual model. Observation wells where residuals were very large (> 5 m) were examined in detail, and compared to other observation wells near to them, the possible range of river water levels if the well was adjacent to the river, ground surface elevation, and the expected water table surface in that area. The RMS error for model was 8%.

Surface Water – Groundwater Interaction

In order to quantify the linkages between the surface water and groundwater regime, a water balance analysis was conducted. The model domain was divided into several water budget zones, and Zone Budget (ZBUD) in Visual MODFLOW was run. In the upper two layers, these zones correspond to the floodplain (extending along the rivers), the various individual irrigation areas, areas not included with the irrigated land (background zone), and the deeper silt and clay layers.

During spring freshet on the Kettle River, the rise in river stage causes inflow of water to various ZBUD zones (after passing through the floodplain area). This excess water is stored in the aquifer. Mass balance calculations indicate that storage rates are less than 50% of inter-zonal groundwater flux, and 15 to 20% of river-aquifer flux. As river stage drops, the hydraulic gradient is reversed; water is released from storage and enters the floodplain zone where it eventually returns to river as baseflow. The rate of inflow to groundwater from the river along the floodplain zone follows very closely the river hydrograph during the rise in river stage. As the river stage levels off and begins to decrease, the flow direction is reversed, generally within 10 days. At this time, the rate of inflow from aquifer to the river begins to rise, and then dominates for the rest of the year, as water previously stored in aquifer drains back to river as baseflow seepage. However, the rates of inflow of groundwater to the river are much smaller, at least an order of magnitude less than the river discharge and the river stage is not affected, as previously discussed. As much of the pumping water is lost to evapotranspiration on irrigated fields, there is a small reduction in the baseflow component to the Kettle River during the pumping period.

The river-aquifer interaction has maximum flow rate of 41 m^3/s, which translates to between 11 and 20% of river flow during spring freshet. Thus, the river puts about 15% of its spring freshet flow into storage in Grand Forks valley aquifer, and within 30 to 60 days most of that water is released back to the river as baseflow. The effect on stage was calculated based on the assumption of no flow loss from the river.

Groundwater flow directions (not shown) are generally downslope in the valley from west to east, and also away from valley slopes and toward the floodplain areas. The flow vectors deviate between river channels, and locally toward pumping wells. Patterns also change seasonally. Vertical groundwater flux (Fig. 11) has a complex pattern in the valley. The positive flux areas (shown in red) along Kettle River in layers 1 and 2 represent outflow of water from the river into the aquifer (influent river reaches). The negative flux areas (shown in blue) are mostly located along the river, and suggest effluent river reaches where aquifer supplies baseflow to river as seepage. River reaches that have inflow or outflow can be identified from these maps.

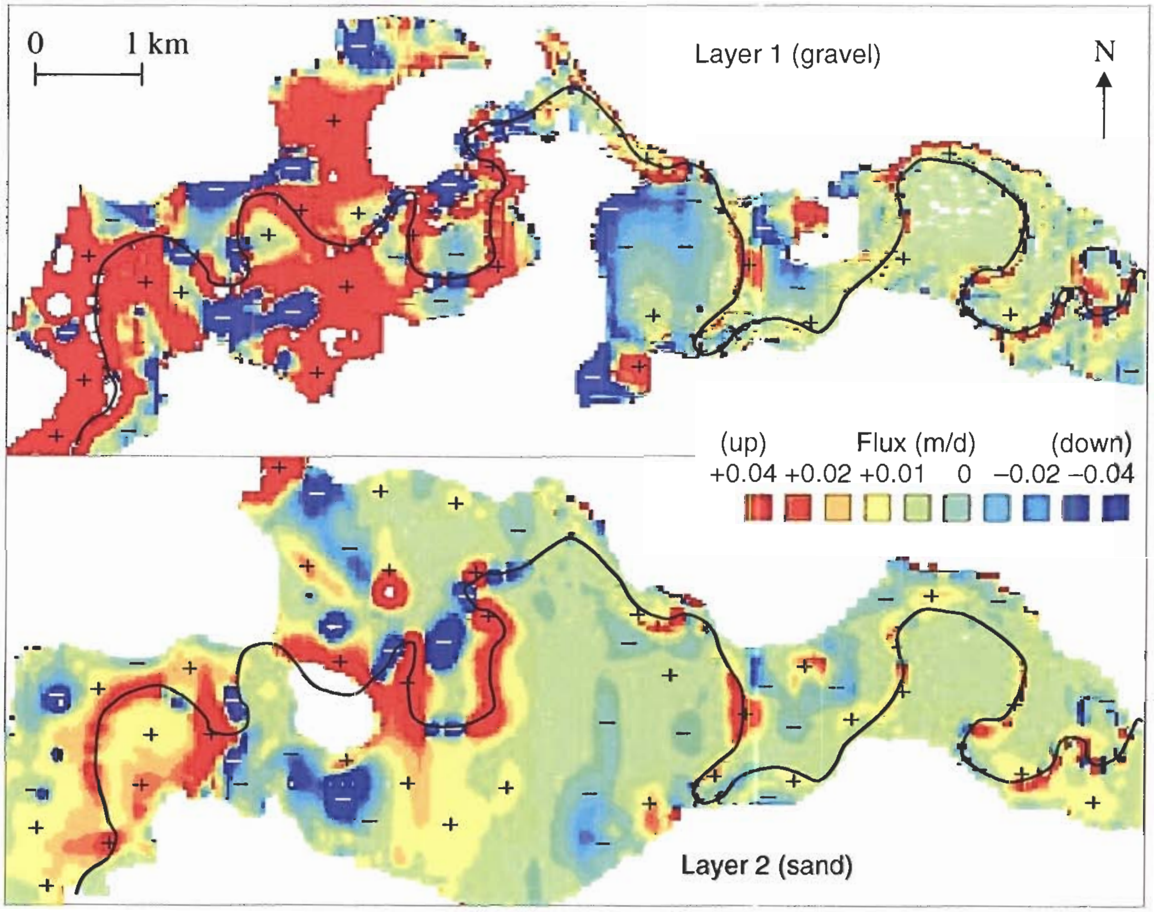

Figure (3) - 11: Groundwater flux between layers shown for top two layers of the groundwater flow model (layer 1: gravel – ground surface; layer 2: sand), as simulated in the model aquifer layers. Location of Kettle River is indicated by black lines on map.

Aquifer Responses to Climate Change

The aquifer responses to modelled climate change are difficult to discern on head distribution maps because the high hydraulic gradient in the Grand Forks valley dominates flow patterns. Climate-induced changes in water elevations are on the order of 0.5 m, while the gradient in the valley spans about 30 m in elevation. Thus, it was necessary to develop a strategy for displaying any changes induced by climate, which would remove the hydraulic gradient of the valley (and valley topography) and allow direct comparison to present conditions. Accordingly, head difference maps (Fig. 12) show only differences due to climate change between future climate scenario model outputs and present climate scenario model outputs (pumping wells activated during summer). Note that because drawdown was identical in all climate scenarios (pumping rates were held constant in all models), ultimately, the pumping effects were subtracted out in these maps. The model responds to pumping exactly as it should and this supports the validity of the model, although we did not try to predict future changes to pumping rates as those are not known.

At day 131, the main cause for the observed changes in head is the shift in river hydrograph peak flow to an earlier date, which creates a positive difference in water levels between the 2010-2039 and 1960-1999 models. In other words, in the future the peak river stage would be earlier and water levels would be higher in the aquifer at an earlier date. The zone of storage is roughly along the river floodplain and also in areas where there are higher river terraces. Within one month, the peak flow passes and river water levels begin to drop rapidly.

By day 160, the river water levels are similar in both the 2010-2039 and modelled present, but only along the river channel. Away from the river channel water levels are elevated by 0.30 to 0.40 m (stored water), which are still draining until day 180. Water levels in the floodplain in the 2010-2039 climate scenario are lower by 0.10 to 0.40 m at day 180 than at present climate at that day, but the temporal shifts in river hydrographs cause changes in aquifer water levels compared to present, when compared to the same day of year. The overall hydrograph shape remains the same, simply shifted earlier in the season. At day 182 the increased recharge in the 2010-2039 climate scenario over historical climate causes up to 0.10 m higher water levels away from the river. The hydrograph shift for the 2040-2069 climate scenario is larger than in the 2010-2039 climate scenario, so the computed differences to historical climate are similarly larger.

Overall, the climate change effects for the 2010-2039 and 2040-2069 climate scenarios relative to present are limited to the floodplain, as well as to the early part of the year when the river hydrograph shifts and is at peak flow levels. As the river peak flow shifts to an earlier date in a year, the aquifer levels (hydrographs) also shift by the same interval, confirming our expectations, but showing a surprisingly strong hydraulic connection between the river and the aquifer.

Impacts of climate change are smallest in those areas least connected to the river (distant from the river and at higher elevations). A small increase (0.1 m) of water levels due to an increase in direct recharge is forecast for future climate scenarios [Scibek and Allen, submitted a], but these increases tend to occur only in areas that are not strongly influenced by the river (i.e., benches at higher elevation around the periphery of the valley). The magnitudes of seasonal groundwater level variations and, locally, pumping drawdown are much larger than climate change effects shown here, but the climate change effects are nonetheless locally significant. One limitation of our approach is that long-term changes in climate were not explicitly modelled, and so our model does not capture the long-term changes in groundwater storage that would result in changes in average static groundwater levels in the valley. Models at this time cannot adequately resolve long-term storage trends (computing constraints) nor manage the uncertainties involved. Notwithstanding these limitations, long-term dynamics should be computed from transient model runs on actual river hydrographs and not averaged ones.

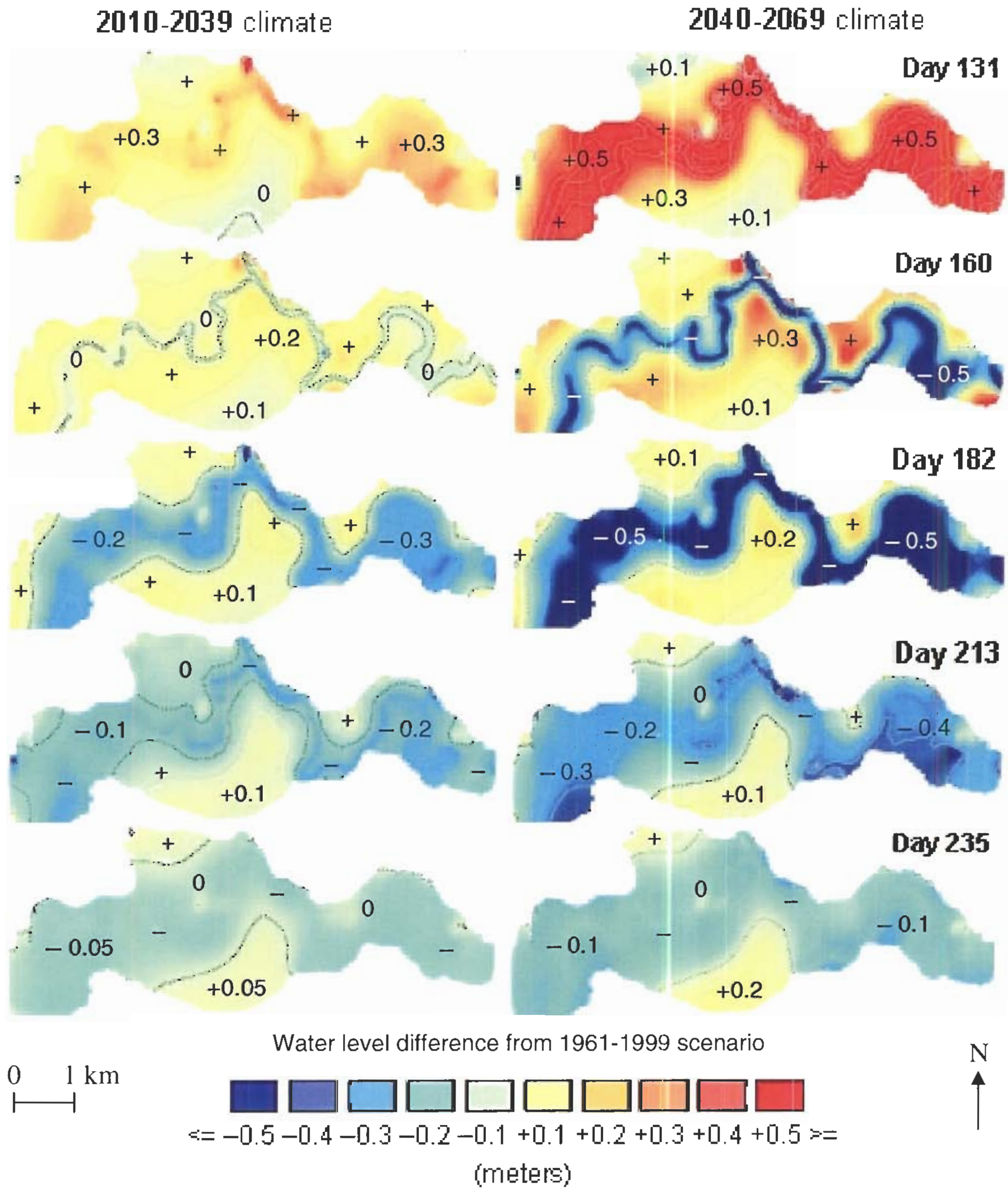

Figure (3) - 12: Water level differences (measured as head in layer 2 of the unconfined aquifer) between a) future (2010-2039) and present climate, and b) future (2040-2069) and present climate under pumping conditions. Maps by time step in days 131 to 180. Positive contours are shown at 0.1 m interval. The zero contour is a dashed line. Negative contours are not shown. Darkest blue colours indicate values < -0.5 m (along rivers only). At day 101 (not shown), difference map has values within 0.1 m of zero. For larger maps see Scibek et al [2004] in Appendix 2.

If the model predictions are correct, and there is a decrease in minimum discharges on Kettle River in late summer and autumn months, a larger percentage of river discharge may be pumped to production wells in the valley (the same rate of groundwater pumping along river but smaller river discharge). The assumption is that the pumping from production wells along the river can induced recharge to aquifer directly from the river water, as supported by groundwater flow model results [Scibek et al, 2004]. Knowing that the river-aquifer interactions are fast, on the order of 30 to 60 days for equilibrating river and groundwater levels in the river floodplain [Scibek et al, 2004], the predicted increase in streamflow (and consequently river stage and groundwater levels in floodplain area of aquifer) in the winter months should not affect the groundwater levels in the following year. However, the shift in peak flow of the river discharge, along with the whole hydrograph curve, will bring lower groundwater levels earlier in the summer.

Conclusions

The water balance analysis and the relation between water levels in the observation well and the Kettle River have established that the unconfined Grand Forks valley aquifer is hydraulically linked to the river. The Kettle River discharge is much greater than the inflow of tributaries in the valley watershed. The river-aquifer interaction has a maximum flow rate between 11 and 20% of river flow during spring freshet – on average, the river contributes about 15% of its spring freshet flow into aquifer storage, and within 30 to 60 days most of that water is released back to the river as baseflow. Storage rates are less than 50% of inter-zonal groundwater flux, and 15 to 20% of river-aquifer flux. Most of the connection between aquifer and river occurs within the floodplain, but under pumping conditions, there is a reduction in the return of water from the aquifer to the river as baseflow compared to non-pumping conditions. The pumping effects in the model are as expected, but the actual changes to pumping rates in the future (e.g. valley development) are not known and were not be used as scenarios.

Future climate scenarios indicate temporal shifts in river hydrographs. These shifts cause changes in aquifer water levels compared to present, when compared on the same day of the year. Modelled water level differences are less than 0.5 m away from floodplain, but can be greater than 0.5 m near the river. However, the overall hydrograph shape remains the same. As the river peak flow shifts to an earlier date in the year, the aquifer water levels shift by the same interval. Impacts are smallest in zones least connected to the river (away from the river and at higher elevation). The hydrograph shift for the 2040-2069 climate is larger than for the projected 2010-2039 climate, therefore the computed differences in water levels for future scenarios compared to historical are similarly larger. The maximum ground water levels associated with the peak hydrograph are very similar to present climate because the peak discharge is not predicted to change, only the timing of the peak.

The groundwater resources in the valley will not be affected significantly by these changes as long as the Kettle River maintains its discharge and supplies large quantities of recharge to the aquifer. Except near pumping wells, the aquifer groundwater levels cannot drop below the Kettle River water levels in the valley, even if there was no direct recharge from precipitation to the aquifer. In the end, the future groundwater use in the valley is limited by the withdrawal of an acceptable percentage of Kettle River discharge, especially at its minimum discharge rate in the late summer.

4 CLIMATE CHANGE IMPACTS ON GROUNDWATER IN THE ABBOTSFORD-SUMAS AQUIFER

Modelled Climate Change Impacts in the Abbotsford-Sumas Aquifer, Central Fraser Lowland of BC, Canada and Washington State, US.

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[Scibek and Allen, 2005 b].

Introduction

Water resources are central to any study on climate change. However, most research to-date has been directed at forecasting the potential impacts to surface water hydrology [e.g., Whitfield et al, 2002]. Relatively little research has been undertaken to determine the sensitivity of groundwater systems to changes in critical input parameters, such as precipitation and runoff. The main reason for studying the interactions between aquifers and the atmosphere is to determine how groundwater resources must be affected by climate variability and climate change. It is expected that changes in temperature and precipitation will alter groundwater recharge to aquifers, causing shifts in water table levels in unconfined aquifers as a first response to climate trends [Changnon et al., 1988; Zektser and Loaiciga, 1993].

This paper describes the results of a climate change impacts modelling study in the Abbotsford-Sumas aquifer. The aquifer is located within the Fraser and Nooksack River Lowlands in the central and eastern Fraser Valley in southwest British Columbia (BC) and northern Washington State (WA) (Figure 1). The aquifer is mostly unconfined and is located on a broad outwash plain, which is elevated above the adjacent river floodplains. The area is also confined by bedrock outcrops. The Sumas Mountain north of Abbotsford is composed of Huntington Formation sandstones and conglomerates of Tertiary age, and older igneous rocks that form the Coast Mountains to the north. Most of the other outcrops to the south, and under the Pleistocene sediment fill in the valley, consist of Tertiary and older sedimentary rocks that form the Georgia Basin, and then much deeper (1 to 3 km) igneous rocks that form the Coast Mountains and Cascade Mountains. To the west, there is a drainage divide along hilly terrain of the township of Langley, BC. The outwash terrace slopes southward, and terminates in escarpments along the Nooksack River floodplain. Small streams drain the area. The aquifer is highly productive, is bisected by the international boundary, and provides water supply for nearly 10,000 people in the US (towns of Sumas, Lynden, and farmlands) and 100,000 in Canada, mostly in City of Abbotsford, but also in township of Langley [Mitchell et al., 2000]. The coastal climate is humid and temperate, with 1000 to 1500 mm mean annual rainfall over most of the year.

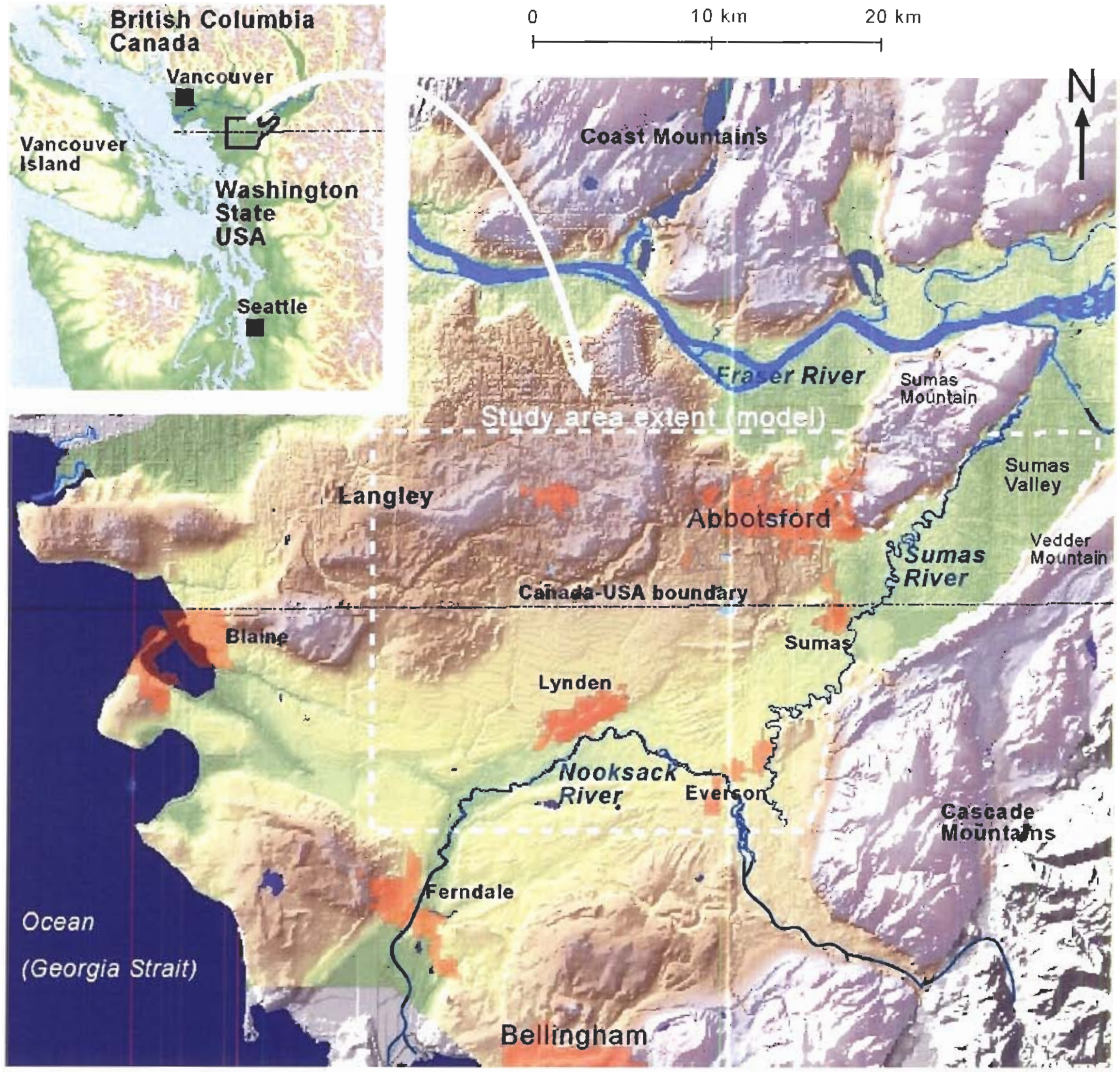


Figure (4) - 1: (a) Regional location map of the model area in British Columbia and Washington State, (b) Central Fraser Valley location map showing model area, cities and towns, topography, international border, and major rivers. White dotted outline shows model boundary, which encompasses the Abbotsford-Sumas aquifer. Urban areas are shown by orange colour on map.

The approach consisted of constructing a three dimensional groundwater flow model for the aquifer, modelling spatially-distributed and temporally-varying recharge, following the methodology of Scibek and Allen [submitted a], based on the historic climate scenario, and then calibrating that model to historic water levels. For the climate scenarios, recharge values for future climate change scenarios were modelled and input into the model, and the impact on water levels in the aquifer calculated. The methodology is consistent with that used by Allen et al. [2004] for the Grand Forks aquifer in south central BC.

Geological Framework and Hydrostratigraphic Model

The following description of the geological framework for the Abbotsford-Sumas aquifer and the Fraser Lowland is summarized from various geological and hydrogeological reports [Clague et al., 1998; Cox and Kahle, 1999; Halstead, 1977; Hunter et al., 1998; Kahle, 1991; Ricketts et al., 1993; Ricketts and Liebscher, 1994] as well as from examination of thousands of borehole lithology logs from drilled water wells in the region.

The Fraser Lowland consists of rolling hills of glacial drift, 60 to 120 m above broad valley floors. The floodplains are currently near sea level. The valley fill consists of complex sequences of diamictons and stratified drift, in various associations with marine and deltaic sediments, showing complex structure and chronology of deposition [Armstrong, 1981]. These sediments had a complex depositional history during the Wisconsin glaciation of the Pleistocene period, during which the lowland experienced repeated glacial and interglacial events, as described in detail in review papers by Armstrong [1981], and Clague [1994].

The Abbotsford-Sumas aquifer is mostly unconfined and is composed of sands and gravels, with dense clay lenses, of the Sumas Drift, a glacial outwash deposit. There is significant heterogeneity of the hydrostratigraphic units, which likely results in complex groundwater paths, particularly at a local scale. The aquifer is underlain by extensive glaciomarine deposits, generally described as glaciomarine stony clays [Armstrong et al., 1965], which are found near ground surface in the Langley area west of the Abbotsford-Sumas aquifer, and interpreted to underlie the surficial aquifers. Laterally, the valley sediments are confined by the Tertiary bedrock surface, which outcrops as mountains on both sides of Sumas Valley, and as small outcrops south of Nooksack River. The elevation of the Tertiary bedrock surface beneath the Pleistocene deposits of the lowland varies considerably, indicating pre-glacial erosional topography with large relief [Easterbrook, 1969]. Near Abbotsford, BC there is about 300 to 500 m of accumulated Pleistocene sediment overlying bedrock. A digital representation of the Tertiary bedrock topography was generated using deep borehole data, existing bedrock contour maps [Hamilton and Ricketts, 1994], valley wall profiles, offshore bathymetric contours, and extrapolated cross-sections through the study area. This surface is considered relatively impermeable and serves as an effective lower boundary to groundwater flow.

Due to the significant heterogeneity of the sediments and the questionable quality of water well records, the traditional approach of constructing cross-sections by interpolating lithologies between boreholes to create a solid model was not possible. Such an approach would invariably lead to a "smoothed and homogenized" representation of the stratigraphy because layers could not be clearly identified. An alternative approach involved examining clusters of boreholes and mapping the lithologies as hydrostratigraphic unit zones (K-zones) directly into MODFLOW. This involved defining, on a layer by layer basis, property zones in Visual MODFLOW [Waterloo Hydrogeologic Inc., 2000] that would correspond to similar hydraulic properties (K and S_s). Geographical Information System (GIS) data visualization allowed conjunctive viewing of borehole lithologs, surficial geology maps, ground and bedrock surfaces, and MODFLOW grid layers (mostly planar surfaces). As MODFLOW requires continuous layers, some judgment was required to create appropriate slice elevations. This was done through GIS, whereby elevation zones were created for each slice surface, then imported to MODFLOW as xyz surface elevation points. Due to the complexity of the stratigraphy, no unique representation was possible. Thus, the mapped geology is analogous to one interpretation of the regional sedimentary structures (similar to one "realization" of a stochastic geologic interpolation in 3D). Notwithstanding, the final representation was based on local geological and hydrogeological interpretations as reported in the published literature. Figure 2 shows the hydrostratigraphic units for layer 3 of the MODFLOW model, and highlights the high degree of heterogeneity that can be captured using this alternative mapping method.

Each K-zone represented in the model was then assigned a unique hydraulic conductivity (K) and specific storage (S_S) value. There is extensive pump test and specific capacity data for the US side of the model, but sparse information on Canadian side. Thus, mean values were calculated from the US dataset for each hydrostratigraphic unit (sampled at well screen location), and the properties extrapolated to areas with poor pump test data. K and S_s data are observed to have a heterogeneous distribution, and strong zonation in some areas (see report by Scibek and Allen [2005 a] in Appendix 4).

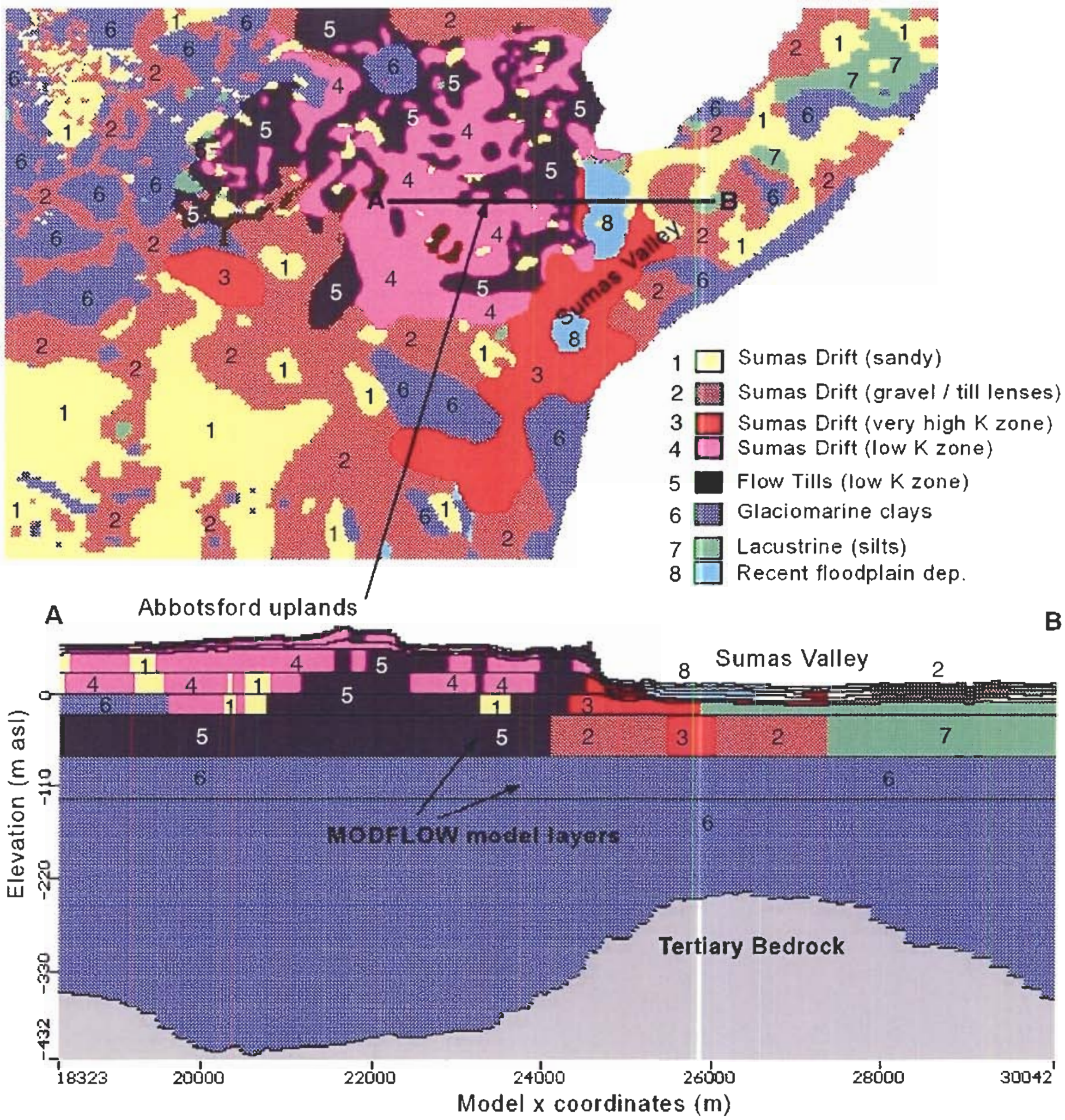


Figure (4) - 2: (a) Map of hydrostratigraphic units in layer three of the MODFLOW model and (b) cross-section from W to E in central region.

The initial model calibration attempts indicated that there were areas with large residuals, which did not respond to changes in K within the reasonable range for each mapped K-zone. In those areas, the geology was re-interpreted from borehole lithologs, this time with much more attention given to possible interpretations, keeping in mind the model residuals, surficial geology, and looking at individual borehole records to verify standardized lithologic units. In many areas, there are many possible interpretations of local geology due to poor distribution of boreholes. The primary problem with the lithology dataset is the uneven distribution of deep boreholes; some areas rely exclusively on interpolated hydrostratigraphic units, which could be interpreted differently. As much care as possible was taken to calibrate this model locally, and to repeatedly review the hydrostratigraphic unit distributions and adjust the hydraulic conductivity values accordingly. The interpretation favouring the lowest possible model residuals was selected, and the geology re-mapped in that area. Therefore, the groundwater flow model provided a feedback to the interpretation of geology in areas with poor distribution or low quality of borehole data.

Climate Scenarios

Climate scenarios for modelled present and future conditions were taken from the Canadian Global Coupled Model (CGCM1) [Flato et al., 2000] for the IPCC IS92a greenhouse gas plus aerosol (GHG+A) transient simulation. Daily data sets for CGCM1 were downloaded from Canadian Institute for Climate Studies (CICS). These include absolute and relative changes in precipitation, including indirect measures of precipitation intensity, dry and wet spell lengths, temperature, and solar radiation. Climate data were downscaled using Statistical Downscaling Model (SDSM) software [Wilby et al., 2002; Yates et al., 2003]. Downscaled data were calibrated to observed historic climate data.

Three year-long climate scenarios were generated using the calibrated downscaled model, each representing one typical year in the present and future (2020s and 2050s):

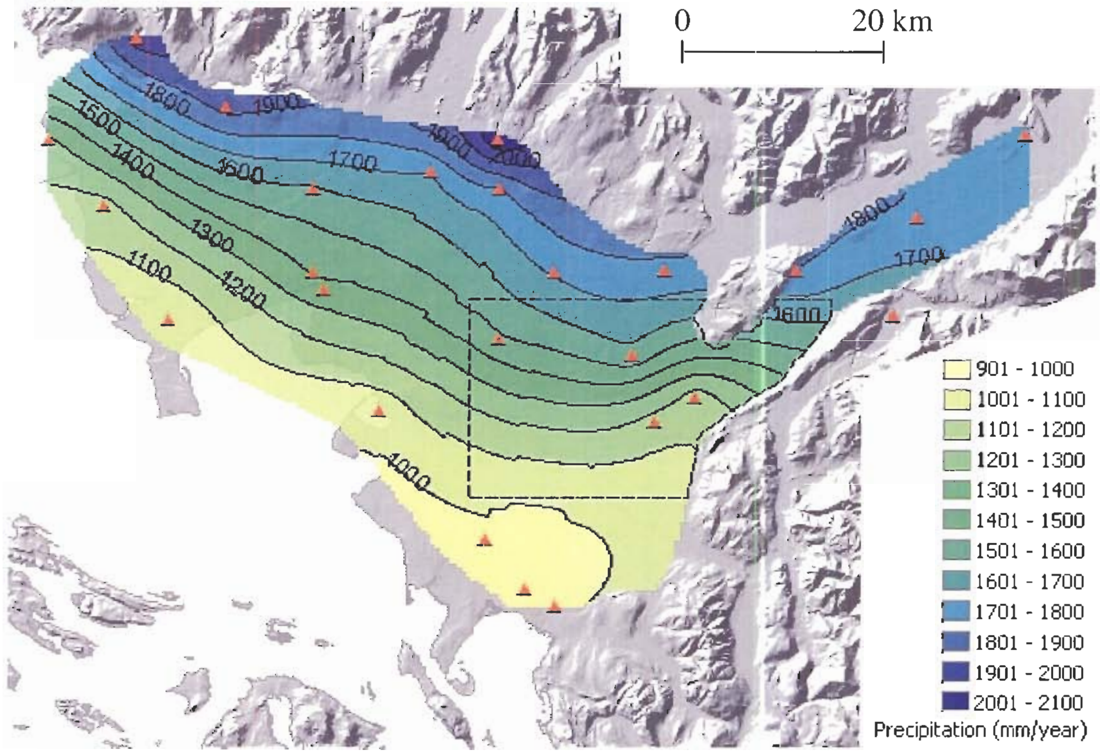
current climate (1961-1999), 2020's climate (2010-2039), and 2050's climate (2040-2069). Daily weather was generated using the LARS-WG stochastic weather generator [Racsko et al., 1991; Semenov et al., 1998]. In this study, only the effects on groundwater levels of changes to recharge are considered.

Recharge Modelling

Aquifer recharge was generated as spatially-distributed and temporally-varying recharge zonation [Scibek and Allen, 2003 b] using GIS linked to the one-dimensional HELP (Hydrologic Evaluation of Landfill Performance). HELP is a hydrologic model developed by USEPA [Schroeder et al., 1994] and the code is contained in UnSat Suite software [Waterloo Hydrogeologic Inc., 2000]. The approach we developed for recharge modelling is similar to that of Jyrkama et al. [2002], in methodology for estimating temporally varying and physically based recharge using HELP for any MODFLOW grid cell. Our method differs from previous distributed-recharge methods in that we also estimate the distribution of vertical saturated hydraulic conductivity in the vadose zone and the thickness of the vadose zone at high spatial resolution [Scibek and Allen, submitted a]. A total of 64 unique recharge zones were defined based on classed soil column properties, and recharge was estimated for each. All map processing was done on 20 m raster grid cells. The temporal inputs are derived from the LARS-WG stochastic weather generator, as opposed to WGEN (internal weather generator in UnSat Suite), and as derived from downscaled CGCM1 predictions.

Recharge estimates were based on soil type, vadose zone property, and rainfall. Mean annual rainfall was subsequently adjusted for the observed precipitation gradient (Figure 3). The precipitation map was computed as percent difference in mean annual precipitation to that recorded at the Abbotsford Airport, which was used as the index station for climate change scenario forecasts. Thus, all recharge estimates were adjusted proportionally by the same percent difference, assuming that recharge is directly proportional to precipitation for any given recharge zone. This is the simplest method of such calculation, otherwise the inputs to the HELP model would have to be estimated for all locations of the model prior to determination of recharge zones by the HELP model output. The overriding assumption is that the precipitation gradient is similar throughout a "typical" year. The gradient magnitudes are different in the 12 months, and the question arises whether the gradient direction is similar to mean annual precipitation gradient. The dominant rainfall volume-wise is frontal and occurs during the winter months, however in the summer the pattern is frequently convective, and may not reflect this pattern. Such processes might account for the seasonal differences in the strength of the precipitation gradient in the valley. In this study, we make an assumption that the gradient in all months equal to the mean annual precipitation gradient, which is approximately true (within 10% of mean annual precipitation in most areas of the valley).

(a)



(b)

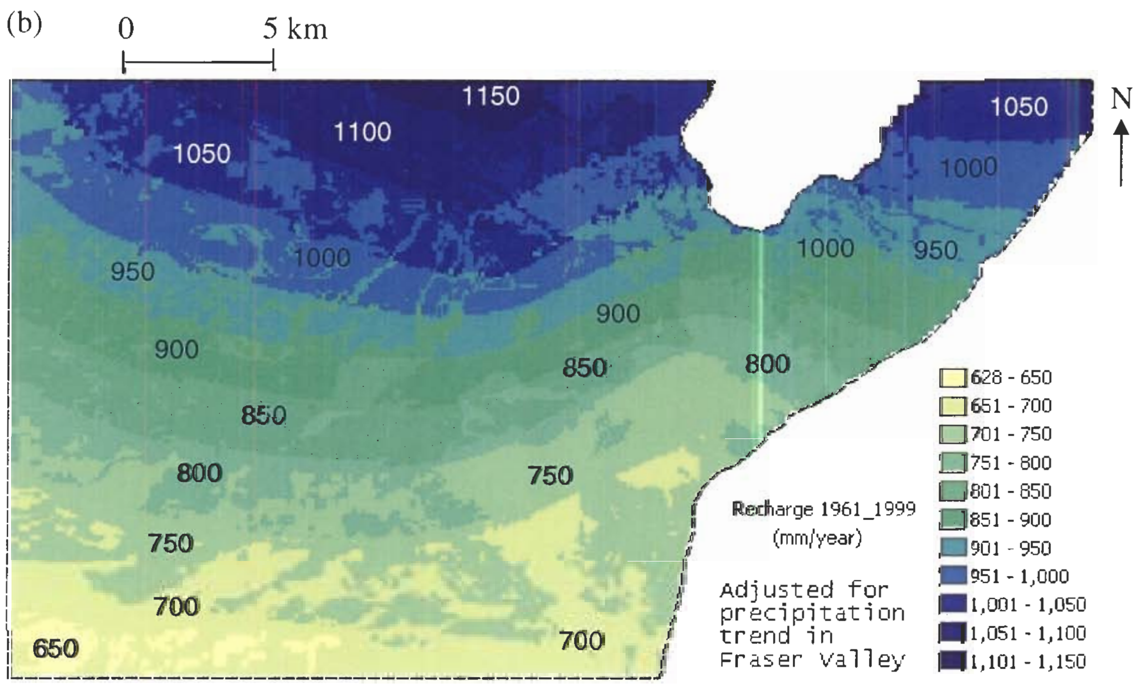


Figure (4) - 3: (a) Mean annual precipitation in the lower Fraser Valley and (b) map of mean annual recharge to aquifer.

Surface Hydrology and Groundwater Interactions

The largest valley in this area is the Sumas Valley, which runs north-east to south-west and contains the lower drainage of the Sumas River (Figure 4). Sumas River flows to the northeast and picks up a significant baseflow component from aquifer discharge on its eastern side. To the south is the Nooksack River, flowing to the west and then south, and draining most of southern drainage. It has baseflow contributions from the Abbotsford-Sumas aquifer, as well as from the aquifers to the south. Most of the surface and groundwater flow from the Abbotsford-Sumas aquifer ends up in the Nooksack River. To the north, the model area includes a portion of the Fraser River floodplain. Several sizable creeks drain to the north, but the quantity of groundwater travelling north is considerably less than that flowing south and west.

The previous investigations in Washington State on streams draining the Abbotsford uplands established that the baseflow component is very high, between 70 and 95% of stream flow in large creeks such as Fishtrap Creek [Sinclair and Pitz, 1999]. Knowing this, and the fact that the stream channels are strongly hydraulically linked with the aquifer, the question arises as to what boundary condition is most appropriate for the groundwater flow model along these streams. In the upper reaches of streams, such as Fishtrap Creek, the stream bed is often perched above the regional water table (Figure 5).

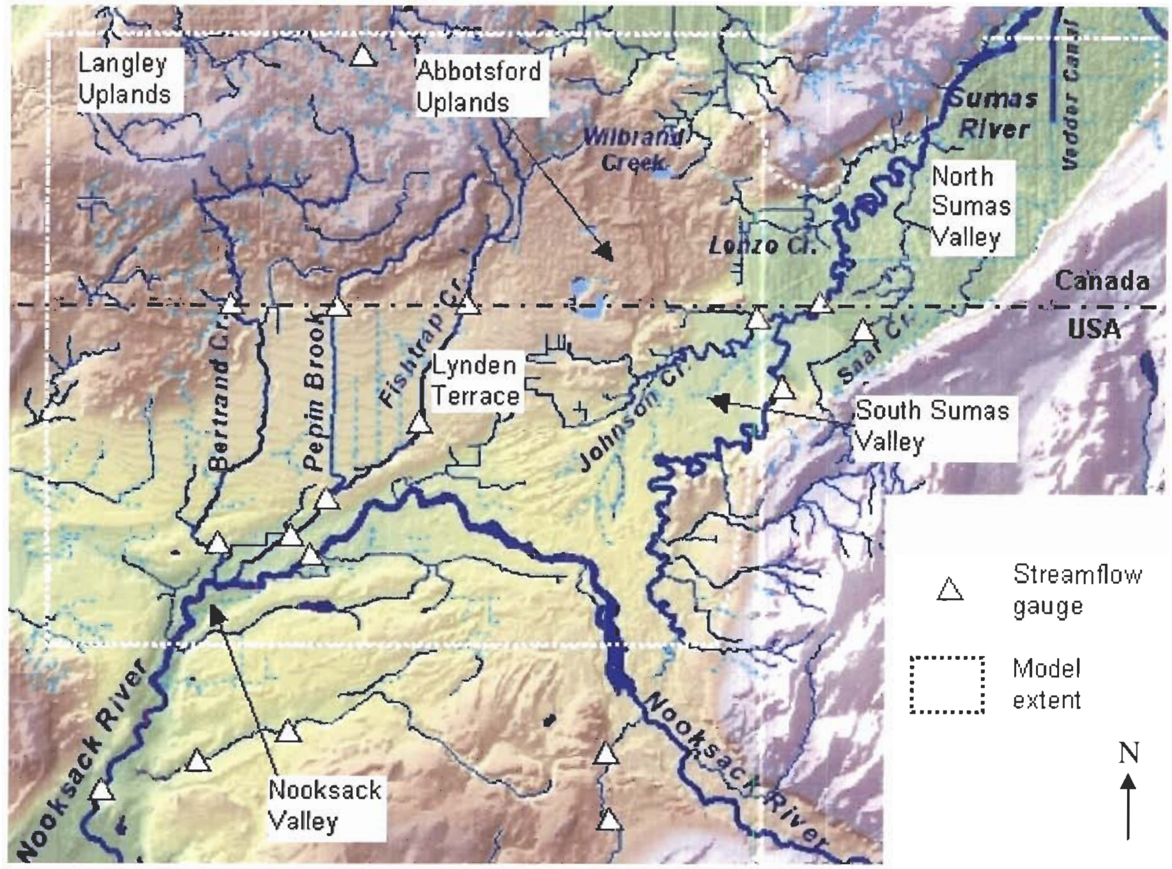


Figure (4) - 4: Streams and rivers of central Fraser Valley, draining the Abbotsford-Sumas aquifer system, and locations of streamflow gauges.

In the lower reaches, the stream receives large inflow from groundwater. Groundwater elevations change by 2 to 4 m seasonally, away from the streams, according to observation well hydrographs [Environment Canada, 2003]. However, stream water elevations vary by much less, although streamflow does change seasonally. Over most of the stream distance, the stream gains groundwater from the aquifer at an average rate of 0.025 m^3/s/km channel length. Thus, it is unlikely that changes in streamflow in a creek such as Fishtrap Creek would affect groundwater elevations in the adjacent aquifer. Therefore, the streams can be represented as specified head boundaries, such that the head schedules will represent the modelled river stage in the transient aquifer model.

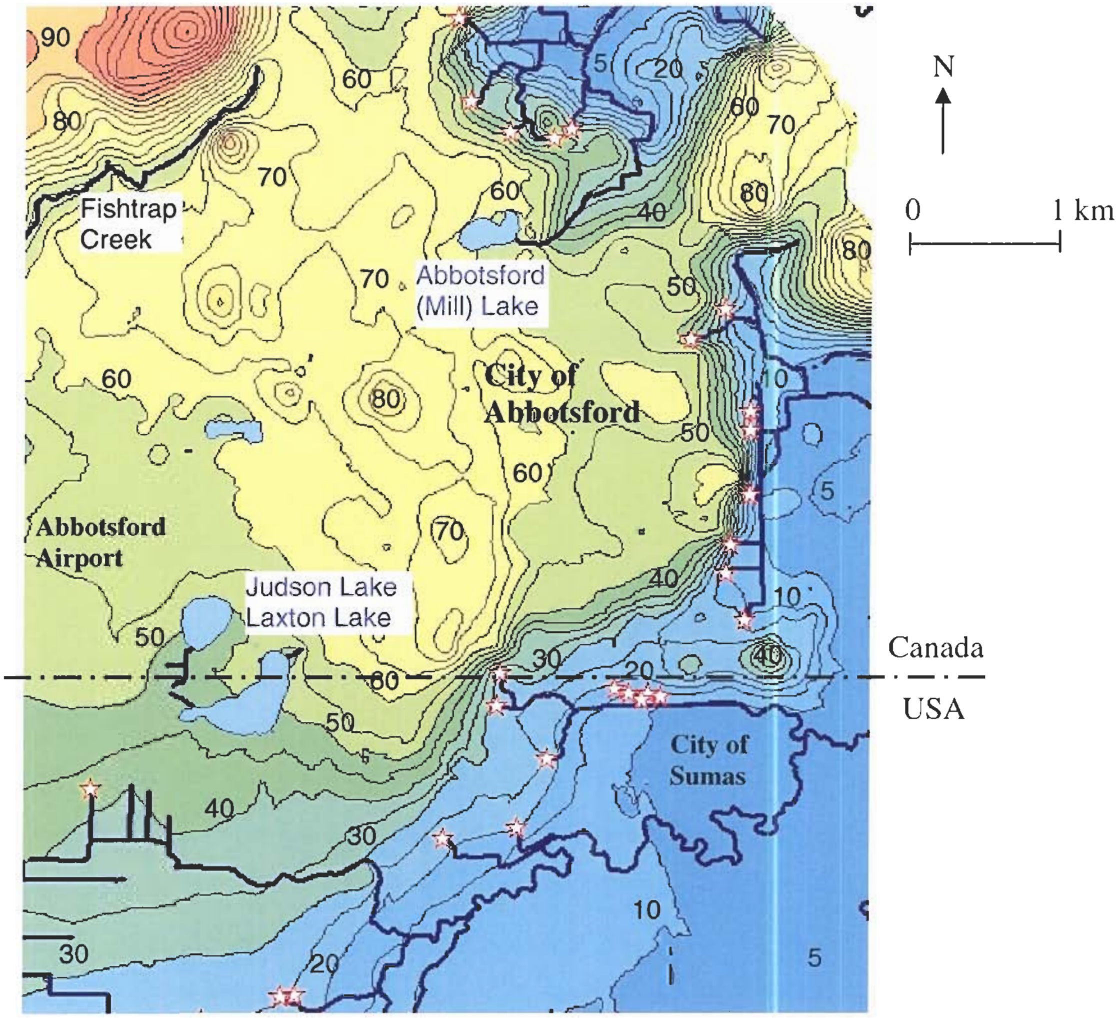


Figure (4) - 5: Locations of lakes, streams, and major springs along scarps of Abbotsford Uplands and Sumas Valley shown with potentiometric surface map, interpolated from available static groundwater and surface water elevations (elevations in meters asl and 5 m contour interval).

The term “constant head” and “specified head” are equivalent here because the head is “constant” for the duration of a time step, but then is specified to change to different value with time. Lakes (Figure 5) that have gauges were modelled as constant (specified) head boundary condition, where a schedule of head values (monthly) could represent the water level in the lake. Other lakes were assigned constant head values for average surface water elevation.

Larger rivers such as Sumas and Nooksack Rivers have seasonally changing discharge and stage hydrographs. However, most of the hydraulic heads in the aquifer above the river floodplains are not affected by changes in river stage. Only the adjacent areas to the river are affected. It is a simplification in the model to represent the larger valley rivers as constant head boundary conditions, without temporally varying stage hydrograph, but as the groundwater flow model covers mostly aquifer area above the valley floodplains and the larger rivers, the assumption of constant head in the larger rivers will not affect model results in those upland areas even in the transient model.

Drain boundary conditions were used for large ditches and ephemeral streams. Drains were used only in areas where the calculated heads were too high above ground (or lake) surface, and drains were used to tie-in the water table elevations to lake and drain elevations.

There are over 2000 groundwater level records in the central Fraser Valley in the study area. The datasets that were selected include all static water levels in the BC well database, all available United States Geological Survey (USGS) and WA Ecology well records, transient water observations from piezometers and observation wells monitored by Environment Canada (south of Abbotsford Airport), USGS, WA Ecology, and others.

A total of 2958 wells with static water levels were used for calibration of the steady-state model. These wells include all of the domestic water wells, of varying depth, and major production wells. The wells are evenly spaced across the aquifer, and thus, provide an excellent means for steady-state and transient model calibration. It is important to recognize, however, that the water elevations used for model calibration were determined at the time of drilling, and, therefore, may not be representative of current groundwater conditions. In this respect, the ability of the model to accurately represent local detail is lower than it would be had the calibration data and stream elevation data been collected at the same time.

Most of the observation wells have very similar temporal variation in groundwater levels. The water table elevation is highest from February to April, then declines in elevation at a non-linear rate until August, when the rate of decline decreases. The minimum groundwater levels occur between September and November. In December, or as early as November, the increased precipitation (in wet years) causes a rise in water table again. At most locations sampled, the amplitudes of the groundwater level hydrographs are between 2 and 3 meters (Figure 6).

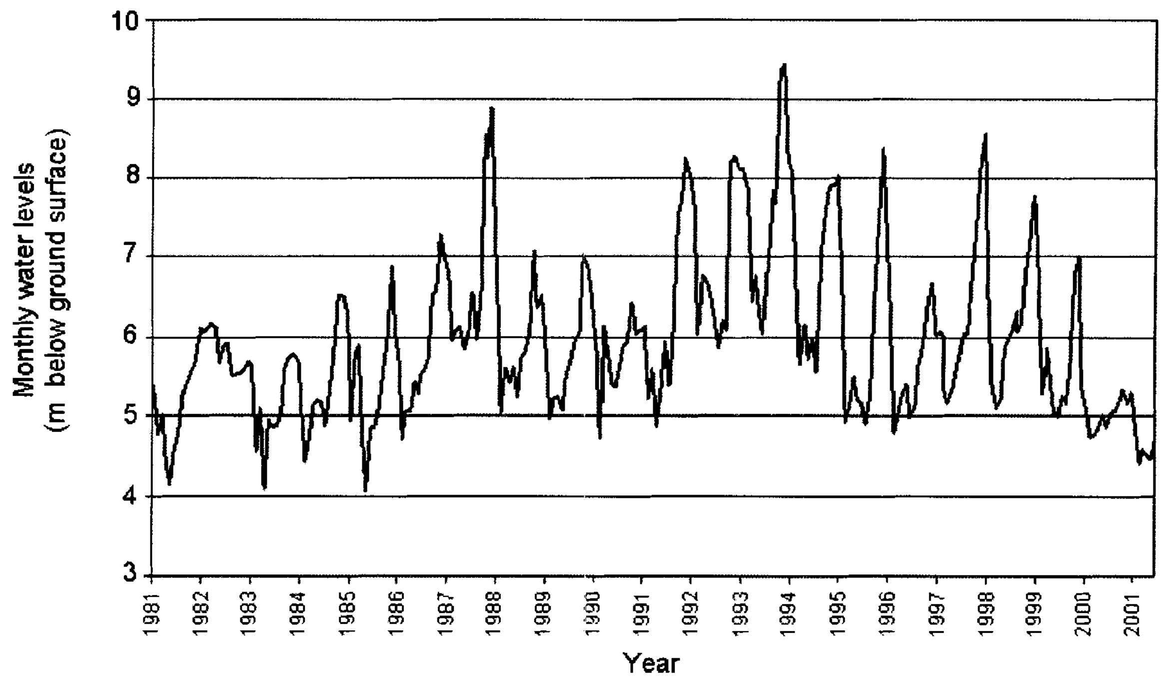


Figure (4) - 6: Monthly water levels for Observation well 272m on Farmer Rd. in Abbotsford showing annual variations in water level on the order of 2-3 m over the period of record (1981-2003). Water levels based on end-of-month readings.

Model Results

The effects of climate change are difficult to observe on head distribution maps because the highly variable and localized hydraulic gradients in the central Fraser Valley dominate all other trends. The climate-induced changes in water elevations are on the order of less than 0.25 m (25 cm) in most areas, but are up to 2 m in sensitive areas in Abbotsford uplands. The water table elevation in the valley ranges from near 0 to above 80 m above sea level (masl) elevation, so any changes cause only a slight shift in the water table contours. Thus, it was necessary to develop a different strategy for displaying any changes induced by climate, which would exclude the hydraulic gradient within the aquifer, and compare directly changes from present conditions. Accordingly, head difference maps were prepared to show only differences due to climate change between future climate scenario model outputs and present climate scenario model outputs. The pumping effects were also subtracted out in these maps because drawdown was identical in all climate scenarios (pumping rates were held constant in all models for the pumping time period).

Instead of using discrete head values at points (wells), the water table elevation map was used for climate change comparisons. The model layer surfaces are very irregular near the ground surface, and the use of the HUF (Hydrologic Unit Flow) package in MODFLOW 2000 and 3D raster-grid approach to hydrostratigraphic unit mapping is not suitable for representing head maps for each model layer separately. In layers 1 to 4, there are large areas with dry cells (no head value available), and only in Layer 5 are there mostly wet cells in the model. Due to the irregular layering of the numerical model grid, the water table lies in layer 1 to 2 in Abbotsford and Langley uplands, then transitions through layer 3 and 4 to layer 5 in Sumas Valley. Head maps would show some confined and unconfined areas blended together.

Impacts of Climate Change on Recharge and Groundwater Levels

Figure 7 shows the predicted changes in aquifer recharge for each of the 2010-2039 and 2040-2069 climate change scenarios. Results are expressed as a percent difference from this historic time period (1961-1999). Both scenarios indicate a reduction in recharge. The 2010-2039 scenario shows a reduction in recharge by 5.6 to 6.3% relative to historic values, while the 2040-2069 scenario shows a reduction in recharge by 12.7 to 14.6% relative to historic.

(a)

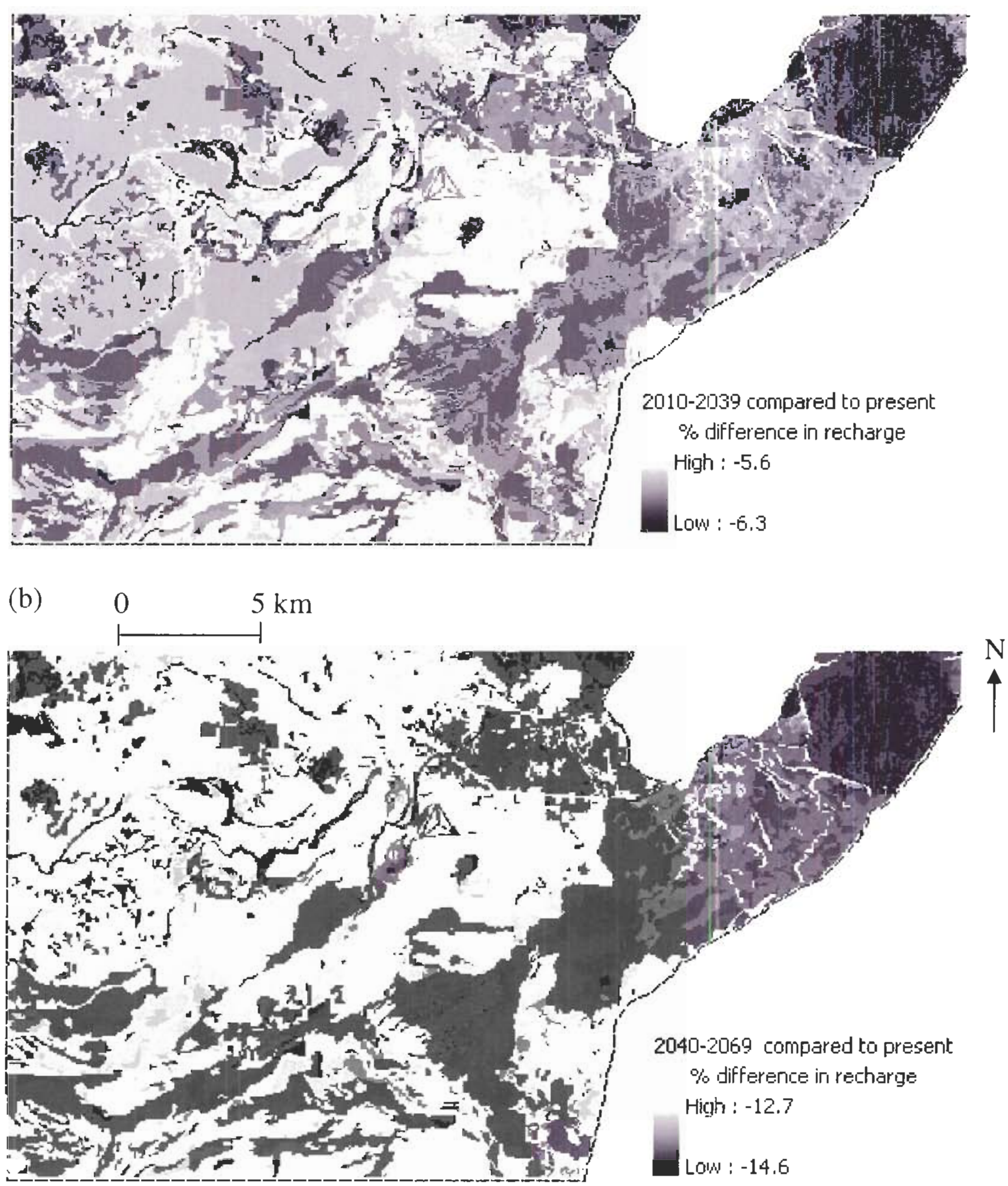


Figure (4) - 7: Predicted changes in recharge to aquifer as percent difference maps from (a) 2010-2039 climate scenario to present, (b) 2040-2069 climate scenario to present.

Figure 8 shows difference in groundwater levels between the future climate scenarios (2010-2039, 2040-2069) and present climate scenario, across the aquifer area, for different days in the transient model. Two trends are apparent for all scenarios at all time steps. First, the water table elevations did not change immediately along river channels where streams, rivers, and lakes were defined as constant head boundary conditions in the model. By definition, constant heads in a flow model do not change. There would be no change expected unless the streams and rivers dried up, or if the timing and magnitude of peak flow changed in the transient model (not simulated here in a transient model). Second, where streams were defined as drains, the water levels were free to vary. This second observation is for areas away from the rivers, where large spatial differences in water level change are observed.

Uplands:

In the Abbotsford uplands, except in a few pockets around lakes and streams, the groundwater levels were predicted to decrease by between 0.05 m to more than 0.25 m due to climate change by the 2010-2039 period. In certain localized areas, such as areas with suspected perched water tables and or poor model calibration, the model predicted very large changes, on the order of 10 m, but those results may be spurious due to poor calibration in areas where there are perched water tables and very heterogeneous hydrostratigraphic units. The decrease in groundwater levels was even greater in the next climate scenario 2040-2069, such that in the Abbotsford uplands, decreases were greater than 0.25 m in most areas. In the Langley uplands, including areas adjacent to the Brookswood aquifer (west), the decreases were smaller in magnitude than in the Abbotsford uplands. In the Langley area, in the 2010-2039 scenario, the groundwater levels dropped by 0.05 to 0.10 m, and in 2040-2069 scenario, dropped by 0.10 to 0.25 m, reflecting the change in percentage recharge.

Lynden Terrace:

The flat and undulating outwash plain north of Lynden, WA, between south flowing Bertrand Creek and Fishtrap Creek, was predicted to have small decreases in water levels (less than 0.10 m). Creeks in that area might be expected to have lower baseflow as a consequence of the predicted lower groundwater levels in the aquifer. Secondly, with lower groundwater levels, the streams could loose water to the aquifer (effluent streams) at certain times of the year due to a reversal in hydraulic gradient. In order to determine the impact of climate change on these streams, it would be necessarily to investigate the existing aquifer-stream connection at different locations, to explore how the interactions might change, using water chemistry studies as supporting evidence.

River Valleys and Floodplains (Lowlands):

The model has excellent calibration (within 1 m) in these areas due to the fact that the valley floor and water table surfaces are flat and because the heads are constrained by the imposed constant head boundary conditions. These are discharge areas of the aquifer, and changes in recharge due to climate change did not produce any noticeable changes in water table elevations in these areas. We may speculate that given such small changes in groundwater table elevations in the future, the future discharge of groundwater into the streams and ditches will be of the same magnitude as present and stream baseflows will not be impacted. However, the actual climate change scenario predictions have large uncertainty and the response of streamflow to changes in groundwater should be monitored, gauged, and surveyed, and the numerical models improved. Nonetheless, the lowering of water table in the uplands would most likely decrease baseflow in those streams fed mostly by the seepage of groundwater.

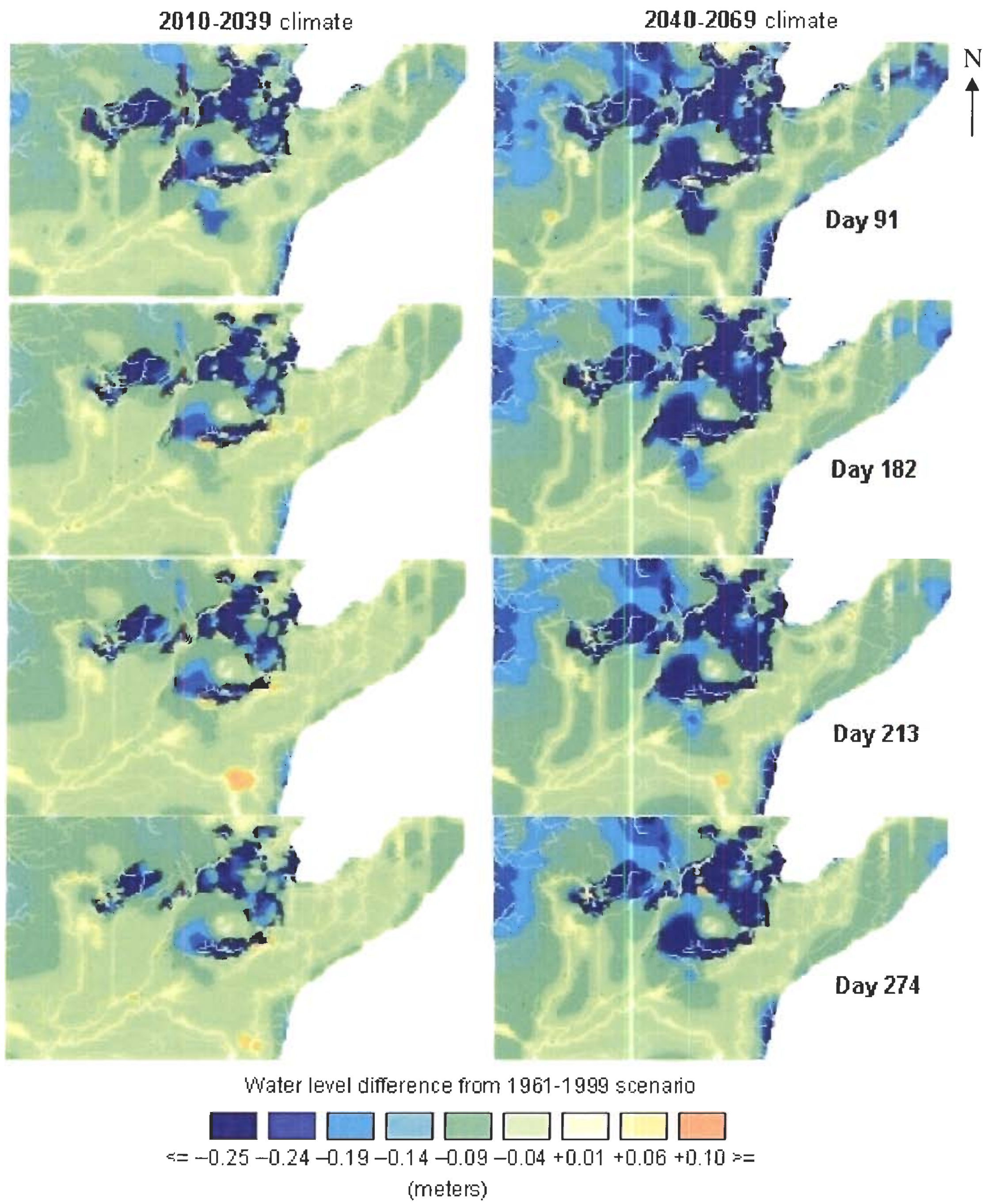


Figure (4) - 8: Water level differences of the modelled water table at days 91, 182, 213, and 274 between future and present climate (a) scenario 2010-2039 and (b) scenario 2040-2069. Values were reclassified to range from 0 to -0.25 m. Values of -0.25 in discrete areas have changes between -0.25 and -3.0 m.

For more details and larger maps see Scibek and Allen [in prep] in Appendix 4.

Conclusions

Recharge to the Abbotsford-Sumas aquifer is predicted to decrease by 5.6 to 6.3% relative to historic values under predicted climate change for the 2010-2039 scenario, resulting in a spatially-variable reduction in water levels ranging from 0.05 m to more than 0.25 m in most upland areas. For the 2040-2069 time periods, recharge is predicted to decrease by 12.7 to 14.6% relative to historic values, resulting in water level declines greater than 0.25 m in most upland areas. These lower water levels will result in a reduction in hydraulic gradients from recharge to discharge areas, and a consequent scaled reduction in groundwater discharge. Because upland streams were assigned as constant head boundary conditions, the model does not predict significant changes in areas adjacent to the streams nor to the streams themselves. The preliminary results suggest that flow rates into streams and ditches are of approximately the same magnitude as observed streamflows, but the lowering of water table in the uplands would most likely decrease baseflow in the streams fed mostly by seepage of groundwater.

In the future climate scenarios, the results of numerical modelling suggest that the aquifer will fill more quickly under increased winter precipitation rates, becoming 'full' some time earlier in the cycle, but under a longer draw down that would accompany longer drier summers that the low water levels might decrease. With the same or a greater amount of annual rainfall falling in a briefer period of time there might also be increased streamflow in winter as the levels in the aquifer might be higher for a longer period, albeit earlier in the year.

Due to the heterogeneity of the aquifer, the nature of interaction between the aquifer and the numerous streams cannot be determined without further detailed investigation. Improvements to the model should consider changes in surface hydrology as a consequence to climate change, but more site-specific information on the streams and refinement of the model in those areas would be needed.

5 A COMPARATIVE STUDY OF CLIMATE CHANGE IMPACTS IN TWO SURFICIAL AQUIFERS

Comparing the Responses of Two High Permeability, Unconfined Aquifers to Predicted Climate Change

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Introduction

With increasing concerns surrounding global climate change, there has been growing interest in the potential impacts to aquifers; however, relatively little research has been undertaken to determine the sensitivity of groundwater systems to changes in critical climate change parameters. It is expected that changes in temperature and precipitation will alter groundwater recharge to aquifers, causing shifts in water table levels in unconfined aquifers as a first response to climate trends [Changnon et al., 1988; Zektser and Loaiciga, 1993]. Precipitation is expected to affect groundwater levels directly, while temperature will have an indirect effect. Where an aquifer is hydraulically connected to surface water, shifts in the hydrologic regime can also be anticipated to impact water levels, although the nature of this interaction may be more difficult to quantify.

Undertaking an assessment of climate change impacts on a groundwater system is complicated because, ultimately, atmospheric change drives hydrologic change, which, in turn, drives hydrogeologic change. For example, groundwater levels near rivers and lakes are directly controlled by changes in surface water levels. Streamflows and lake levels are driven by runoff amounts from precipitation and snowmelt, which are directly controlled by atmospheric conditions. Local precipitation on the aquifer extent also controls directly recharge to surficial aquifers, but regional precipitation and temperature indirectly control groundwater levels as well, through streamflows and water levels in surface water. Precipitation frequency and intensity should also affect recharge rates to surficial aquifers, especially in the summer months when evapotranspiration rates are high. Climatic change may affect any or all of the above, and also rates of groundwater withdrawal (e.g. irrigation). Therefore, actual predictions of climate change impacts on groundwater involve a very thorough modelling effort.

The assessment of groundwater conditions requires detailed information about the subsurface; information that is traditionally difficult to obtain. Each aquifer is unique in its physical properties (geology), it geometry (controls on the hydraulic gradient), and the nature of the connection with surface water bodies. Thus, each aquifer requires specific characterization; often quantification (e.g., numerical modelling) is needed in order to determine what the potential impacts of climate change might be.

In our study, we developed new and improved existing methods for linking climate change predictions to regional scale aquifers. To accomplish realistic links between regional climate, station-specific climate, and the groundwater system at an appropriate scale, we selected two small regional aquifers (less than 150 km^2 in area) to test high resolution groundwater flow models, climatic inputs through recharge, and climate-driven surface water links.

The general approach consisted of constructing a three-dimensional groundwater flow model for each aquifer (using Visual MODFLOW version 3.1.84, [Waterloo Hydrogeologic Inc., 2004]), modelling spatially-distributed and temporally-varying recharge (annual variation) based on the historic climate, applying that recharge to the groundwater model, and then calibrating it to historic water levels. For the climate scenarios, recharge values for future climate change scenarios were modelled and input into the calibrated model, and the response of water levels in the aquifer quantified. This approach assumes that homeostasis exists, that the models calibrated to observed climate and water levels is appropriate for modelling future scenarios and the model parameters are not affected by the forecast changes in climate – that climate controls only recharge to the aquifer.

Climate Scenarios

Climate scenarios for modelled present and future conditions were taken from the Canadian Global Coupled Model (CGCM1) [Flato et al., 2000] for the IPCC (Intergovernmental Panel for Climate Change) IS92a greenhouse gas plus aerosol (GHG+A) transient simulation. Daily data sets for CGCM1 were downloaded from Canadian Institute for Climate Studies (CICS) website. These include absolute and relative changes in precipitation, including indirect measures of precipitation intensity, dry and wet spell lengths, temperature, and solar radiation. Climate data were downscaled using Statistical Downscaling Model (SDSM) software [Wilby et al., 2002; Yates et al., 2003], and downscaled data were calibrated to observed historic climate data. Note that a second downscaling method based on principal component analysis was also used (PCA K-nn). At this particular study location, neither SDSM nor K-nn adequately models precipitation. The performance of the two downscaling models was described in Scibek and Allen [submitted a]. If the critical recharge period to the aquifer is from February to June, also corresponding with peak streamflow in the Kettle River during spring months, the SDSM model has better fit to observed precipitation in that time period than the K-nn model. Details concerning downscaling and the comparison of methods are provided in Allen et al. [2004]. Three year-long climate scenarios were generated using the calibrated downscaled model, each representing one typical year in the present and future (2020s and 2050s): current climate (1961-1999), 2020's climate (2010-2039), and 2050's climate (2040-2069). For recharge modelling, daily weather was generated using the LARS-WG stochastic weather generator [Semenov et al., 1998].

The Study Sites

The Abbotsford-Sumas (AB-SUM) aquifer is located in southwest British Columbia (BC) and northern Washington State (WA) (Fig. 1). It covers an area of approximately 150 km^2 within the Fraser and Nooksack River lowlands in the central and eastern Fraser Valley. The aquifer is highly productive in water yields, is bisected by the international boundary, and provides water supply for nearly 10,000 people in the US, and 100,000 in Canada, mostly in the City of Abbotsford. The coastal climate is humid and temperate, and has strongly seasonal rainfall with a distinct winter heavy rain season and a summer less rainy season.

The valley fill consists of complex sequences of diamictons and stratified drift, in various associations with marine and deltaic sediments, showing complex structure and chronology of deposition [Armstrong, 1981]. Laterally, the valley sediments are confined at depth by the Tertiary bedrock surface, which outcrops to the north and south of the aquifer study area, and by the older Cascade Mountains to the east. The bedrock surface is considered relatively impermeable and serves as an effective lower boundary to groundwater flow. The Abbotsford-Sumas aquifer itself is mostly unconfined and is composed of un-compacted sands and gravels of the Sumas Drift, a glacial outwash deposit of late Pleistocene age (Fraser Glaciation). There is significant heterogeneity of the hydrostratigraphic units, which likely results in complex groundwater paths, particularly at a local scale, as suggested by previous groundwater investigations [e.g., Cox and Kahle, 1999], the analysis of over 2500 borehole lithology logs, and numerical modelling results [Scibek and Allen, 2005 a]. The aquifer is underlain by an extensive glaciomarine deposit, which outcrops to the north and northwest, forming a thick low-permeability confining unit in that area (dark grey-brown area near Langley in Fig. 1). The thickness of the surficial aquifers (drift deposits) ranges from about 20 m to 100 m, whereas the total thickness of Pleistocene deposits is up to 500 m in the central parts of the study area (although much of this thickness is comprised of the glaciomarine sediments).

The Grand Forks aquifer (GF) is located within a small valley in the mountainous and relatively dry climate of the south-central BC interior (Fig. 2). The Kettle River meanders through the valley. The aquifer covers an area of 34 km^2 along the border between BC and WA, and is surrounded by metamorphic rock. The stratigraphic

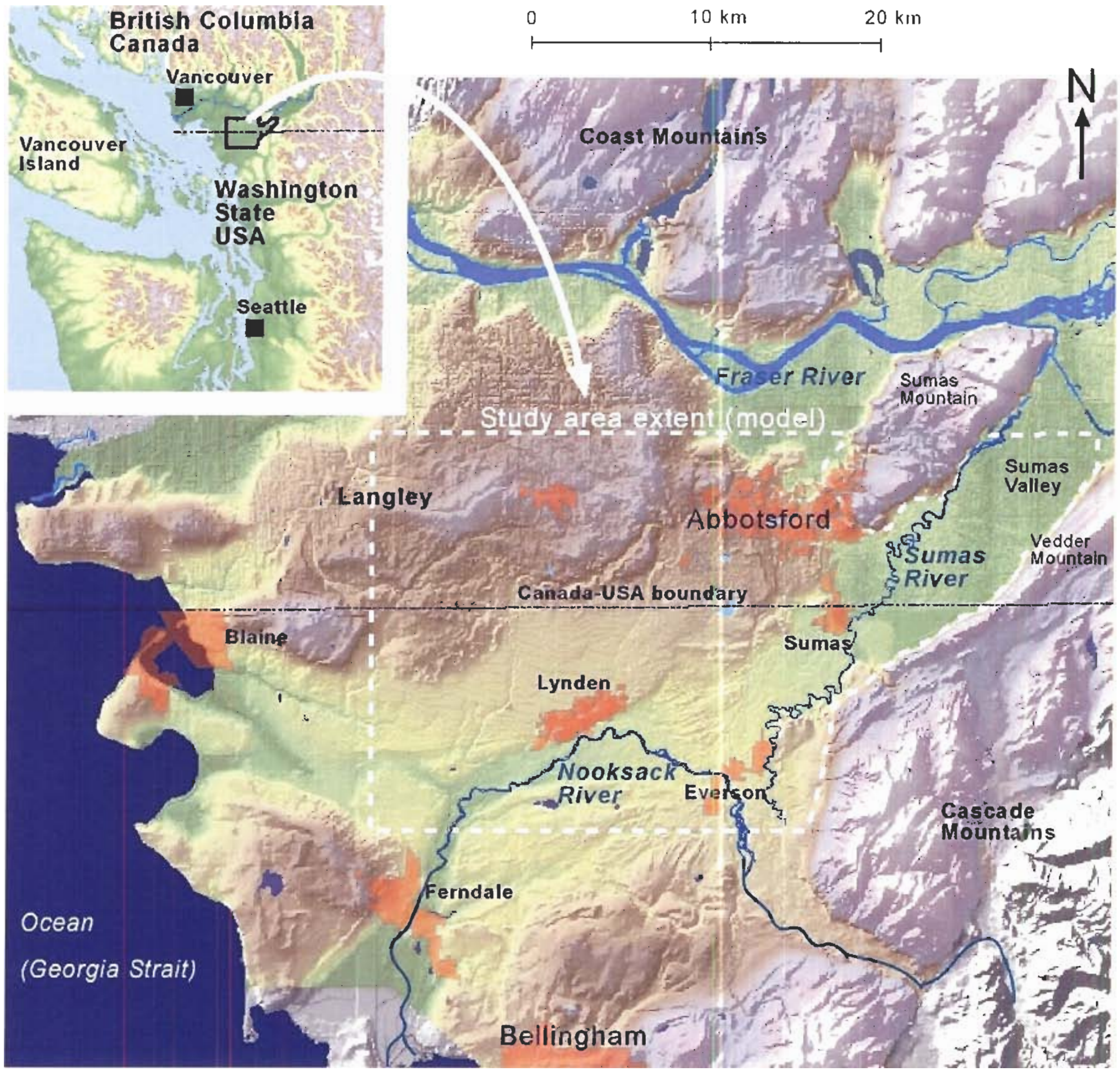


Figure (5) - 1: Abbotsford-Sumas (AB-SUM) aquifer location in British Columbia, Canada and Washington State, USA. Inset map at top left shows location of the study area in a more regional context. The white dashed line shows the extent of the model domain. The orange coloured areas indicate urban areas, while the green-brown range indicates topography, except grey areas of bedrock outcrops and mountains.

sequences at depth are poorly understood; approximately 150 shallow groundwater well lithology logs are available. The upper stratigraphic unit of the aquifer consists of fluvial gravel, overlying sands and gravels. Deeper units are glaciolacustrine silts and, most likely, tills. Sediment thickness in the valley does not exceed 100 m in most places.

Groundwater flow occurs mostly in a surficial gravel unit, which is strongly hydraulically connected to the river, as evidenced by the aquifer water balance, and the hourly synchronous relation of water levels between an observation well in the aquifer and the Kettle River.

Groundwater Model Construction

To construct each groundwater flow model, first, the valley shape was modelled using profile extrapolation, constrained by well lithology logs, and geostatistical interpolation. The hydrostratigraphy was interpreted from selected high-quality well lithology logs, with layering constrained by the Quaternary depositional history of the valley sediments. Hydrostratigraphic units were modelled in three-dimensions from standardized, reclassified, and interpreted well borehole lithology logs. Solid models were constructed using GMS (Groundwater Modelling System) software (v. 4.0) [Brigham Young University, 2002], converted to a five layer system underlain by solid bedrock, and imported into Visual MODFLOW [WHI, 2004], as is typically done with complex multi-layer aquifer systems [Herzog et al., 2003]. Details of the GF model construction are described in Allen et al. [2004], and for the AB-SUM aquifer in Scibek and Allen [2005 a]. Representative homogeneous and isotropic hydraulic properties were assigned to each layer, based initially on values determined from pump test data, but later adjusted during model calibration. The models were calibrated to replicate the observed static groundwater levels as well as the temporal variation in water levels, where such data were available (discussed later).

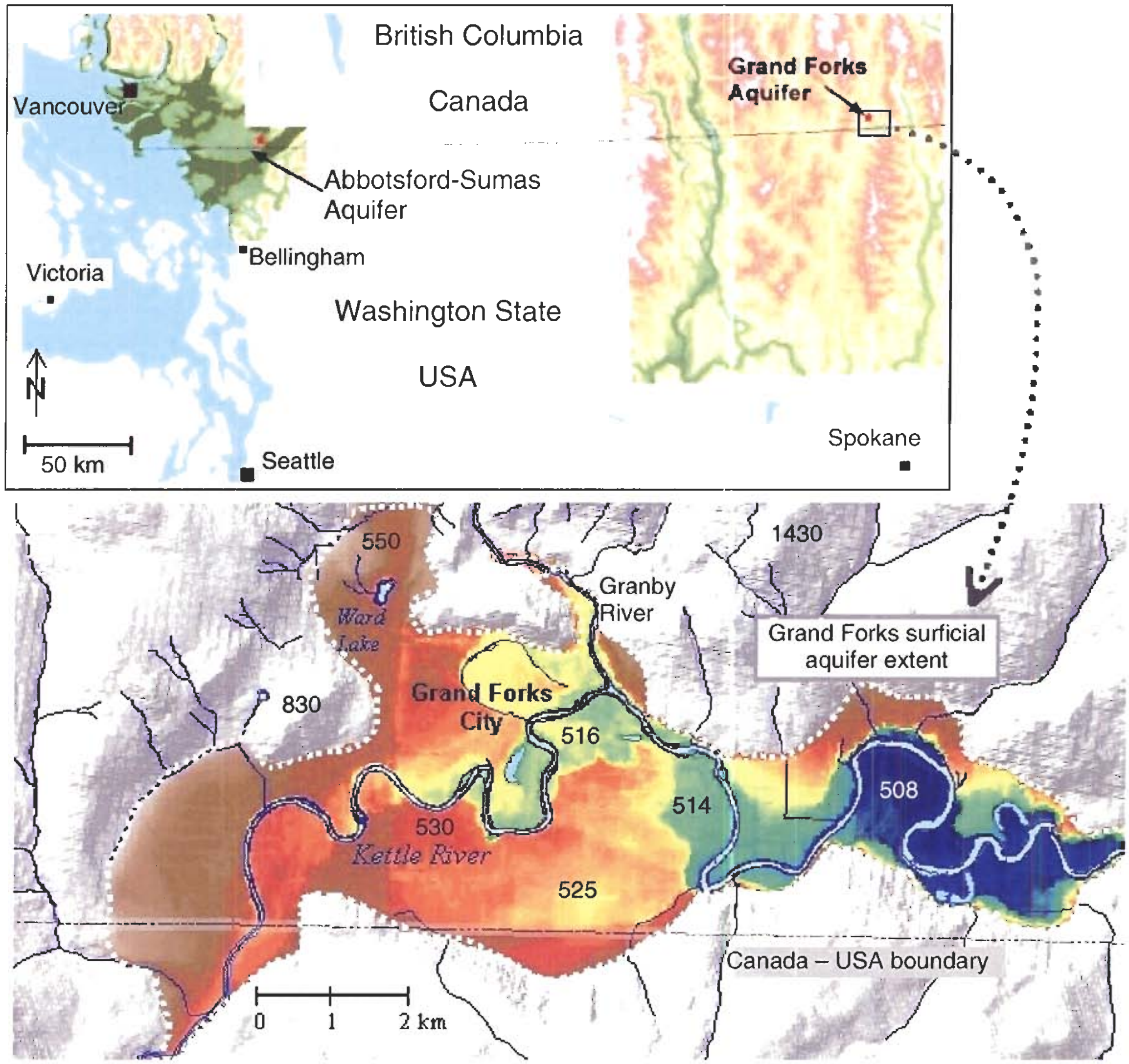


Figure (5) - 2: Grand Forks (GF) aquifer location in British Columbia, Canada. Top figure shows the relative location of the AB-SUM aquifer. Colour shading for GF aquifer indicates topography, high (brown) to low (blue), showing spot elevations. Bedrock areas are grey.

Hydrology

In both study areas, surface bodies, such as streams are represented with model grid cells of 5 to 10 m width where possible. Tens of kilometres of streams and ditches were included in the groundwater flow models using GIS. In the AB-SUM aquifer, runoff is roughly one third of estimated precipitation over the catchment area [Connely et al., 2002]. The Sumas and Nooksack rivers draining the valleys around the AB-SUM aquifer receive a significant baseflow component from the aquifer. The largest valley in the AB-SUM region is the Sumas Valley, which west drains from the south-west to the north-east and contains the lower drainage of the Sumas River (Fig. 1). Sumas River adds a significant baseflow component from aquifer discharge on its eastern side. To the south is the Nooksack River, flowing to the west and then south. The Nooksack has significant contributions from the Abbotsford-Sumas aquifer and from aquifers to the south [Sinclair and Pitz, 1999; Connely et al, 2002]. Many of the small streams (e.g., Bertrand Creek, Pepin Creek and Fishtrap Creek) flow from the uplands southward. To the north, the modelled area includes a portion of the Fraser River floodplain.

Previous investigations in Washington of streams draining the Abbotsford uplands (e.g., Fishtrap Creek) established that the baseflow component is very high, between 70 and 95% of stream flow [Sinclair and Pitz, 1999]. In the upper reaches of such streams, the stream bed is often perched above the regional water table, determined from our water level survey data, static water table elevation map (this study), and model results. In the lower reaches, the stream receives large inflow from groundwater [Sinclair and Pitz, 1999]. Away from the streams, groundwater elevations change by 2 to 4 m seasonally according to observation well hydrographs. Stream stage varies by much less, although streamflow varies seasonally [Hii and Liebscher., 1999].

Within the groundwater flow model, small lakes, swamps, ephemeral streams, and the upper reaches of the streams, are represented with drain boundary conditions. Drains only affect the flow model when water table rises to or above the drain elevation. The drain then takes groundwater out of the aquifer to simulate seepage and baseflow. In effect, the model can be calibrated to simulate filling of drains during high water table levels (high recharge months) and to dry the drains during low water table levels (low recharge months). Drain nodes were used only in areas where the flow model calculated heads too high above ground (or lake) surface to compared to actual groundwater levels. The drains are not assumed to be in contact all year with saturated zone of the aquifer, and many are probably perched above mean annual water table elevation. There is simply a lack of data to verify which of these drains are linked to the aquifer, and to what extent.

Flowing streams are represented in the groundwater flow model by specified head boundary conditions, with the stream channel profiles represented as accurately as possible. With this type of boundary condition, the river can leak and receive water to and from the aquifer, but the river stage will not change as a result of such interaction. In other words, the river will act as an inexhaustible supply or sink of water and will influence the aquifer water levels, but the aquifer will not have any effect on river discharge and stage. It is a simplification in the model to represent these larger valley streams as constant head boundary conditions, but because the groundwater flow model covers mostly aquifer area above the valley floodplains that contain the larger rivers, the assumption of specified head will not affect model results in those upland areas, even in a transient model. Nonetheless, should one want to examine potential changes in streamflow caused by stream-aquifer interactions and feedbacks from climate change, these boundary conditions could be modified to perhaps drain boundary conditions, bearing in mind that model convergence in the absence of more rigorous boundary condition constraints (e.g., specified heads) could be problematic.

In the GF valley, Kettle River discharge is significantly greater than the inflow of tributaries in the valley watershed [Allen et al., 2004]. In most years, at low flow in August, the Kettle River maintains a discharge of between 10 and 14 m^3/s, compared to a minimum discharge of 0.0137 m^3/s from the creeks and baseflow from the aquifer. Thus, the combined aquifer and tributary contribution to the rivers have very small effect on Kettle and Granby River water levels. In contrast, the river water levels have a strong effect on groundwater levels in the aquifer, and the bottom sediments of the Kettle and Granby Rivers above the GF aquifer consist of mostly gravels, with very little fine sediments. Therefore, the rivers are represented as specified head boundaries, such that the head schedules will represent the modelled river stage in transient groundwater flow model.

Recharge

Aquifer recharge was modelled as spatially-distributed and temporally-varying recharge zonation [Allen et al., 2004] using GIS linked to the one-dimensional HELP model (Hydrologic Evaluation of Landfill Performance) [Schroeder et al., 1994]. Recharge estimates were based on soil type and depth, vadose zone conductivity, and water table depth. The approach used for recharge modelling is similar to that of Jyrkama et al. [2002], who proposed a methodology for estimating temporally-varying and physically-based recharge for any MODFLOW grid cell. Our method differs from previous distributed-recharge methods in that we also estimate the distribution of vertical saturated hydraulic conductivity in the vadose zone, and the thickness of the vadose zone, at high spatial resolution. The temporal inputs are derived from the LARS-WG stochastic weather generator at daily time steps [Semenov and Barrow, 1997]. Recharge is calculated daily and, later, averaged monthly for use in transient groundwater flow models.

In the Fraser Valley, the site of AB-SUM aquifer, the mean annual rainfall was represented as a precipitation gradient (1000 to 1600 mm/year) [Environment Canada, 2004; Western Regional Climate Center, 2004] interpolated from high quality climate normals (Fig. 3a). To link the precipitation gradient over the valley to daily precipitation records at the Abbotsford Airport weather station, which was used as an index station for climate downscaling, the percent difference in mean annual precipitation to that recorded at the Abbotsford Airport was calculated. Thus, all recharge estimates were adjusted proportionally by the same percent difference, assuming that recharge is directly proportional to precipitation for any given recharge zone. The overriding assumption is that the precipitation gradient is similar in all months throughout a "typical" year. The modelled recharge for the AB-SUM aquifer is between 650 and 1150 mm/year, and is controlled mostly by the local precipitation gradient. The dominant rainfall volume-wise is frontal and occurs during the winter months, however in the summer the pattern is frequently convective, and may not reflect this pattern. Such processes might account for the seasonal differences in the strength of the precipitation gradient in the valley. In this study, we make an assumption that the gradient in all months equal to the mean annual precipitation gradient, which is approximately true.

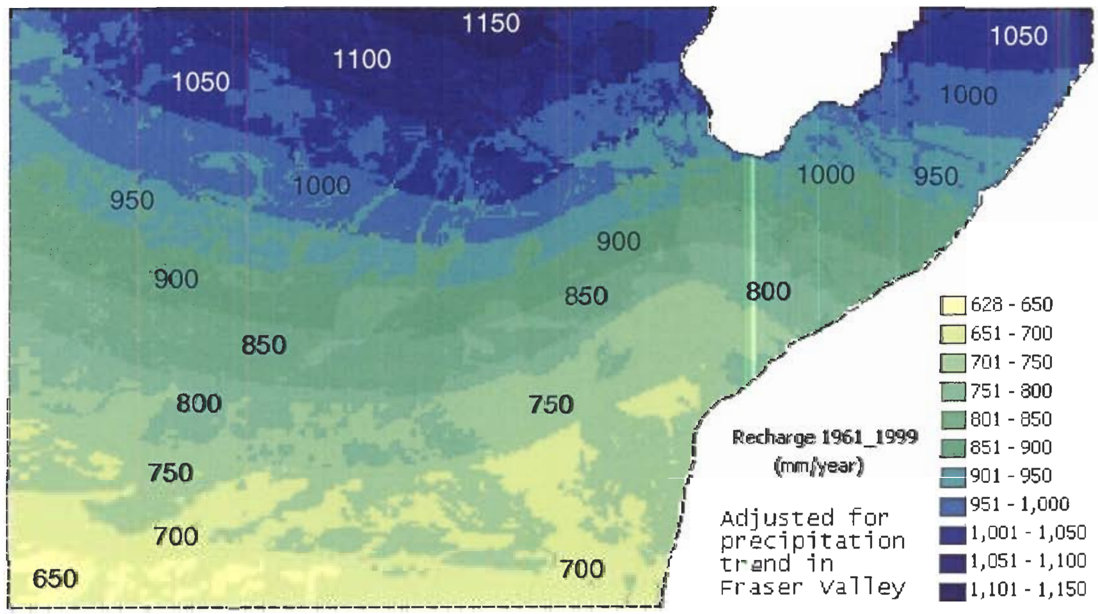
In the GF model, a uniform rainfall distribution was assumed. The modelled recharge at GF was between 30 and 120 mm/year [Scibek and Allen, submitted a], depending on location (Fig. 3b). In this semi-arid climatic region, there is in sufficient precipitation to recharge the aquifer where there are thick sand and gravel terraces – most of the precipitation changes moisture content in these areas, but little of it recharges the groundwater aquifer. The depth of the unsaturated zone is less important in the AB-SUM aquifer where large amounts of rainfall infiltrate and where moisture content remains high.

The AB-SUM groundwater system is mostly controlled by aerial recharge, and the maximum groundwater elevations are constrained by topography and local surface drainage. The AB-SUM aquifer system is much more complex than the GF aquifer, as it includes many perched water table areas and heterogeneous porous media (the ranges in magnitude of recharge between the two sites is from <30->120mm/year in Gf and >628 - <1150 mm/year in AB-SUM (Figs. 3a and 3b). At both study sites, the soil and other properties of the subsurface control the spatial variation in recharge (up to 100 mm/year), but the two locations are very different in recharge rates. The question is does the spatial heterogeneity have a large effect on recharge in the high precipitation or low precipitation area.

Predicted Climatic Change

For both study aquifers the air temperatures are predicted to increase in all months from present to future. After downscaling of CGCM1 climate predictions, we noted that the summer temperatures will increase at a relatively constant rate of 1°C per 30 years, up 3°C by end of century compared to present. In other seasons, the increase can reach up to 4 to 6°C by 2080s, at a relatively constant rate of increase from present [Allen et al., 2004].

(a)



(b)

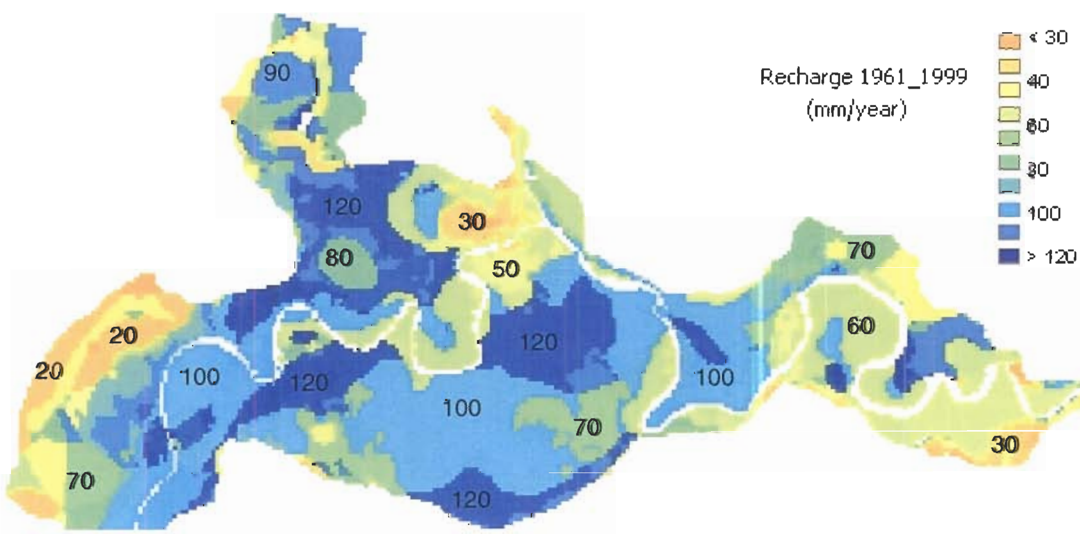


Figure (5) - 3: Map of mean annual recharge to the (a) AB-SUM aquifer, (b) GF aquifer.

Summer precipitation was much more difficult to model from downscaled GCM runs than winter precipitation. For the interior of BC, at the location of the GF aquifer, there is a serious limitation of using CGMC1 predictions. CGCM1 is unable to adequately model precipitation in the summer months (convective activity, high intensity rainfall, rain-shadow effects), giving an underestimate of rainfall of up to 40% compared to observed, even after downscaling with a well-calibrated model, because the observation station is in the valley bottom, and the GCM is for a regional grid cell and does not predict rainfall per se. This is referred to as the "model bias", where the bias is between the downscaled global climate model and the observed. At the coastal location, containing the AB-SUM aquifer, there is a smaller model bias for summer months (than at GF location), and the CGCM1 downscaled climate data matches observed historic data reasonably well. Model calibration to winter precipitation has smaller model bias for both locations, but the downscaling models are better calibrated (smaller deviations from monthly mean precipitation compared to observed) for the AB-SUM coastal location than in the interior, mountainous location of GF.

In the GF aquifer model, we also obtained predictions for basin-scale runoff and discharge in the Kettle River, which strongly interacts with the aquifer at the study site [Allen, 2001; Allen et al, 2003]. The runoff model was derived from a statistical downscaling method (PCA K-nn), which links river discharge to GCM predicted climate [Cannon and Whitfield, 2002]. River stage and water elevation were calculated from stage-discharge curves along 67 channel cross-sections, interpolated along river length, and then input to the groundwater flow model as boundary conditions, using the method described in Chapter 2. The predictions for future climatic conditions indicate that the river hydrograph peak will shift to an earlier date, although the peak flow magnitude will remain the same as at present (Fig. 4). This shift is consistent with type of shift in response measured over the past two decades [Whitfield and Cannon, 2000b]. In the winter months there will also be increased streamflow caused by more late rain events and an increase in snowmelt related to predicted temperature increases in southern BC. Greater streamflows in the winter may contribute to greater recharge to the aquifer. Changes to the river hydrograph are predicted to be much larger by 2040-2069 than in the period 2010-2039, and compared to the modelled 1961-1999 time period. The AB-SUM aquifer differs from the GF aquifer in that a significant portion of the aquifer lies in the uplands, above the floodplains of larger rivers, which may also experience hydrograph shifts as a result of climate change. But, there groundwater levels would only be affected in small areas of the aquifer adjacent to the rivers.

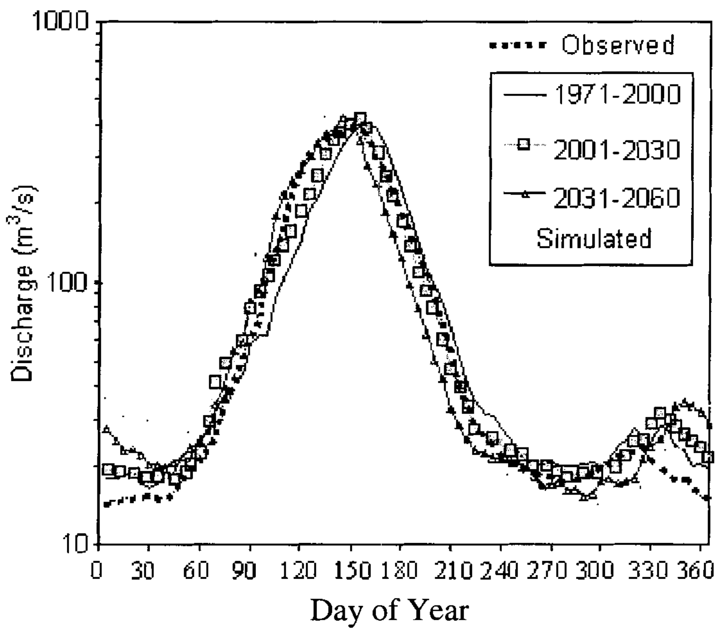


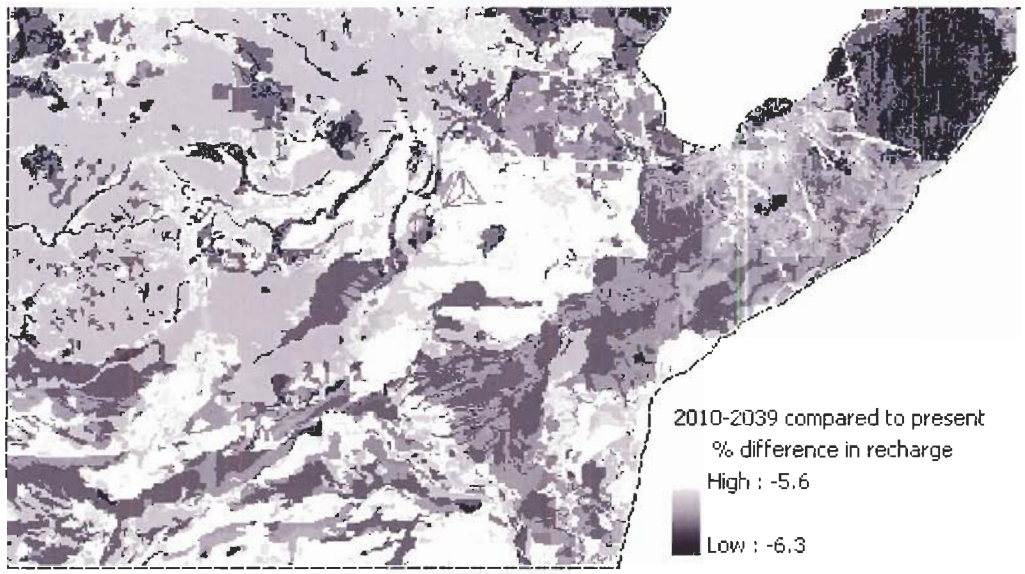
Figure (5) - 4: Predicted discharge in Kettle River at Laurier (WA) modelled using statistical downscaling model and comparing to observed discharge in last 30 years.

The predicted changes in recharge to the aquifers are presented in Figure 5. The changes are depicted as percent change in recharge from modelled present to the 2010-2039 climate scenario. At the AB-SUM aquifer, recharge is predicted to decrease between 5.6 and 6.3 % compared to present, with variation on the order of 1% attributed to heterogeneity of soils and aquifer media (Fig. 5a). However, for the GF aquifer (Fig. 5b), recharge is predicted to increase by between 2 and 6 %, depending on location in the valley. At GF, much of the recharge occurs in the spring during snow melt but also in the summer during convective rain events. At this time, the climatologic research does not indicate whether such rain events may change in intensity or frequency of rainfall. In contrast, at the Abbotsford-Sumas aquifer, most of the recharge occurs as rainfall during the winter months [Environment Canada, 2004].

Model Calibration

Despite a high density of data model calibration for the AB-SUM aquifer was difficult. These difficulties include: presence of perched water tables, spatial clustering of well lithology and water level data (usually along major roads and in areas with higher population density), conflicting data in the databases (including problems with units and terminology in Canadian and US databases), strong heterogeneity of the sediments, and an uncertain hydraulic conductivity distribution in some areas. The normalized RMS was 7.15% using roughly 1700 static water levels from drilled wells. The transient model predicted roughly the observed 2 to 3 m seasonal variation in groundwater levels. However, it was not exceptionally well calibrated for all locations for transient conditions, due to poorly defined three-dimensional distribution of hydraulic conductivity and storage properties. In the river valleys and floodplains (lowlands), however, model calibration was excellent (within 1 m of observed), based on the distribution of residuals between observed and modelled static groundwater levels.

(a)



(b)

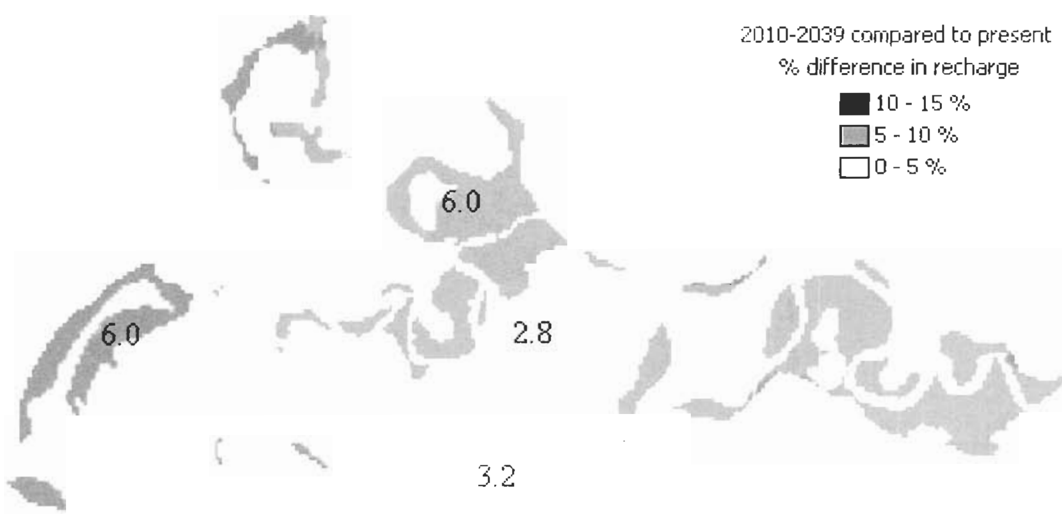


Figure (5) - 5: Predicted changes in recharge to the (a) AB-SUM aquifer, (b) GF aquifer, as percent difference maps from 2010-2039 climate scenario to modelled present.

Note that colour scale and legend is different for each of the two maps and colour scale was selected to show highest contrast of values on each map separately.

In the GF aquifer, there were many fewer static water levels (roughly 300), with a less than an ideal spatial distribution, but the inclusion of detailed river water levels helped to constrain the model to observed groundwater levels through well-defined boundary conditions. The normalized RMS was 8.29%.

Impacts on Groundwater Levels

At both aquifers the effects of climate change are difficult to distinguish on head distribution maps because the highly variable and localized hydraulic gradients in the aquifers dominate all other trends. Impacts of climate change on water levels were better represented by head difference maps for different model time periods. The differences were calculated between the output from each future climate scenario model and the output from the present climate scenario model.

<u>AB-SUM Aquifer</u>:

In the Abbotsford uplands, the main recharge area of the surficial AB-SUM aquifer, the groundwater levels were predicted to decrease by between -0.05 m to more than -0.25 m due to climate change by the 2010-2039 period [Scibek and Allen, 2005 b] as shown in Figure 6. The decrease in groundwater levels was even greater in the next climate scenario 2040-2069, such that in the Abbotsford uplands, groundwater level decreases were between -0.10 and -0.25 m in most areas. The models did not predict any increases in water table elevation resulting from climate change, only decreases, and the magnitude of the decrease depended on location in the aquifer. In places with suspected perched water tables, the uplands under city of Abbotsford and westward (Figure 1), which tended to be areas of poor model calibration, the changes were between -0.5 and -3.0 m.

As a consequence of reduced groundwater levels, streams in upland areas, which were treated as drains, are expected to have lower seasonal flows. In lowland areas, creeks that drain the AB-SUM aquifer did not produce any significant changes in water table elevation due to changes in discharge (Fig. 6). This result is not surprising, given that both the valley floor and the water table surface are generally flat, and are constrained in the model by constant head boundary conditions. What we expect to see, under a regime of lower recharge, and resulting lower groundwater levels, is a shift in the nature of the groundwater-surface water dynamics for entire streams or stream reaches. Streams at lower elevation could become perched above the water table at certain times of the year, particularly during intense rainfall events, thereby loosing more water along their channels and contributing to indirect groundwater recharge (i.e., becoming effluent streams rather than influent streams). A more likely consequence of reduced groundwater levels across the aquifer would be a lowering of the hydraulic gradients, and a consequent reduction in stream baseflow, particularly during the summer months as less groundwater is released from storage. To investigate the complex nature of the interactions between groundwater and surface water in this aquifer, a coupled groundwater-surface water model should be used, and consideration should be given to shifts in the hydrologic regime of all streams, as was done for Kettle River in the GF aquifer as discussed below.

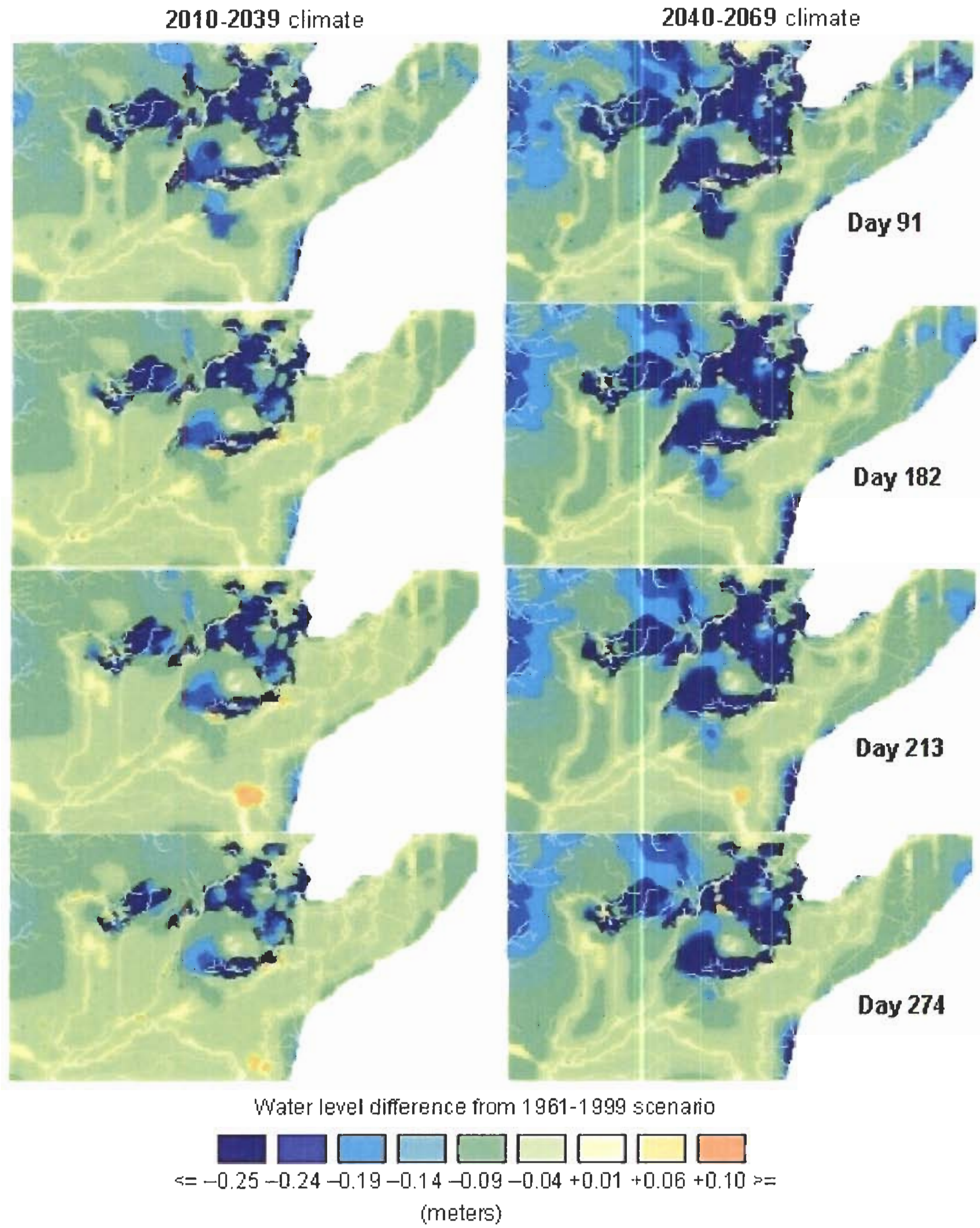


Figure (5) - 6: Water level differences of the modelled water table at days 91, 182, 213, and 274 between future and present climate (a) scenario 2010-2039 and (b) scenario 2040-2069. Values were reclassified to range from 0 to -0.25 m. Values of -0.25 in discrete areas have changes between -0.25 and -3.0 m.

For more details and larger maps see Scibek and Allen [in prep] in Appendix 4.

GF Aquifer:

In this aquifer, the effects of changing recharge on groundwater levels are very small compared to changes in the timing of basin-scale snowmelt events in the Kettle River and the subsequent shift in the hydrograph [Allen et al., 2004]. During spring freshet in the Kettle River under current conditions, the rise in river stage causes an inflow of water to the aquifer, where it is stored for 30 to 60 days. As river stage drops, the hydraulic gradient is reversed, and water is released back to the river as baseflow. The river-aquifer interaction has a maximum flow rate equivalent to 11-20% of Kettle River flow during spring freshet, suggesting that the relationship of river and aquifer stages will be unchanged.

In the 2010-2039 scenario, groundwater levels rise and fall with the river hydrograph earlier in the year relative to historic conditions, because of the shift in peak river flow to an earlier date. When comparing groundwater levels at the same Day-of-Year (the difference in water level from the climate change scenario compared to present as shown in Fig. 7), elevated water levels, up to 0.30 to 0.40 m, persist along the channel into the early summer months. From late summer to the end of the year, water levels near the river are generally lower than at present as a result of both the timing of the flow and a small reduction in streamflow during late summer (see Fig. 4). Away from the floodplain, groundwater levels are similar to present conditions, with small increases observed due to the increase in recharge.

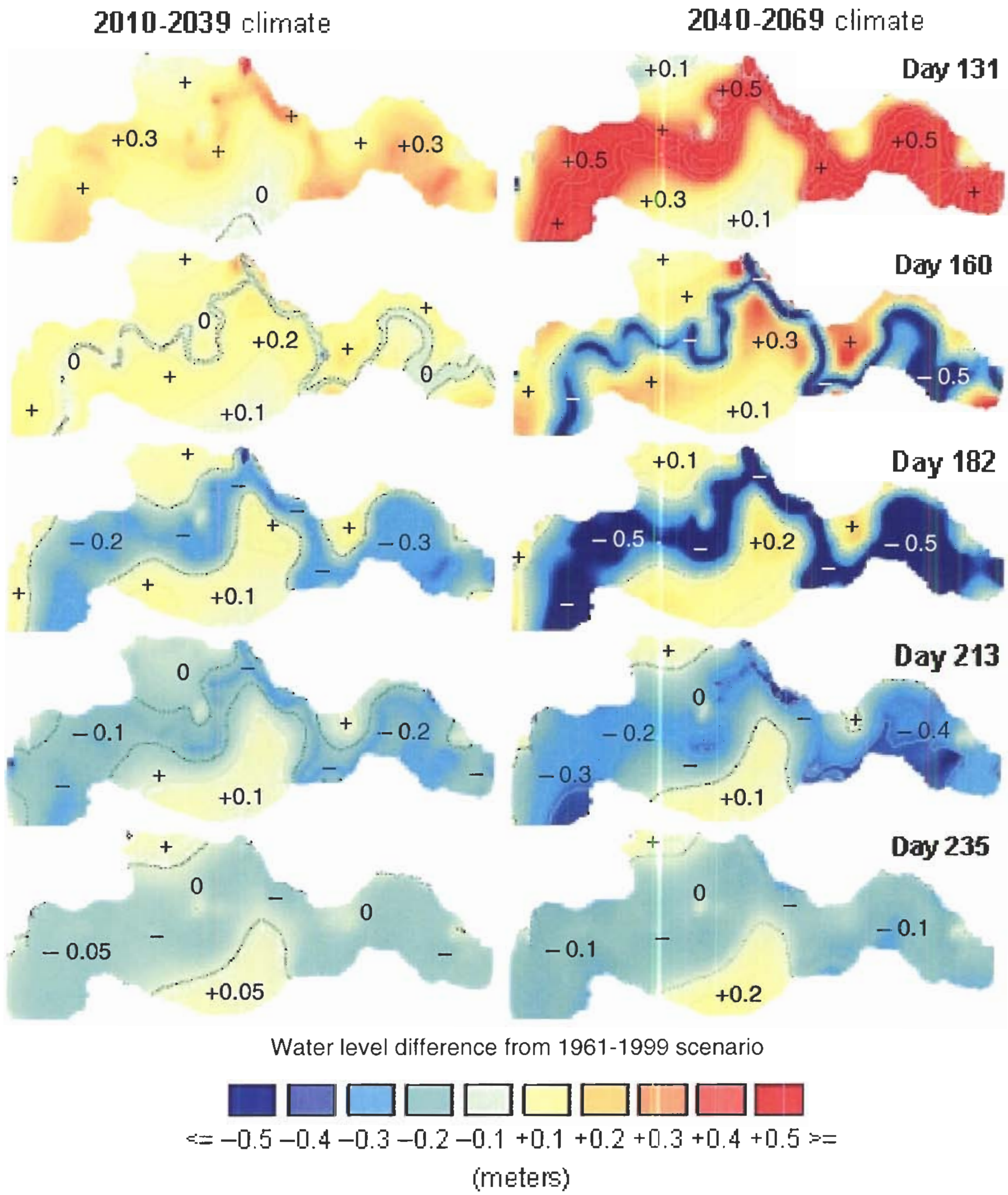


Figure (5) - 7: Water level differences of the modelled water table at days 131, 160, 180, 205, and 235 between future and present climate (a) scenario 2010-2039 and (b) scenario 2040-2069. Positive contours shown at 0.1 m interval. Zero contour is dashed line. Negative contours not shown. Darkest blue colours indicate values < -0.5 m (along rivers only). At day 101, difference map (not shown) has values within 0.1 m of zero.

For larger maps see Scibek et al [2004] in Appendix 2.

In the 2040-2069 climate scenario, groundwater levels are higher by up to 0.50 m near the river and up to 0.20 m away from the river, as a result of the even greater shift in timing of peak flow compared to the 2010-2039 scenario. Changes in recharge from precipitation remain minor in importance.

Conclusions

The climate-to-groundwater model link and associated methodology was successfully applied to two separate small regional aquifers. The downscaled CGCM1 climate predictions were linked through local weather stations to recharge models for the aquifers. Summer precipitation was much more difficult to model from downscaled CGCM1 runs than winter precipitation, especially at the GF location, due to the inability of the downscaling process to match the observed precipitation from the CGCM1 climate model outputs. Some of the reasons for poor downscaling results are mountainous location of the site, rain shadow effects, and local convective rain events that are not clearly related to this scale climatic variables as modelled in the CGCM1 model. At the coastal location (AB-SUM aquifer), there is much smaller model bias for summer months and the CGCM1 downscaled climate matched observed reasonably well. Winter precipitation had smaller model bias for both locations than for the summer months, but it was still better at the coastal location than in the interior location.

Overall, the groundwater flow models allow to discern relatively small impacts of changes in climate on the groundwater systems studied. In the rainfall recharge-dominated AB-SUM aquifer, groundwater levels are predicted to decrease by between 0.05 m to more than 0.25 m due to climate change by the 2010-2039 period. Impacts on water levels are generally restricted to the upland areas, because the lower elevation portions of the aquifer, where the major streams are located, are constrained by specified head boundary conditions; although, reductions in baseflow are anticipated due to the lowering of the groundwater gradient across the aquifer. In the GF aquifer, climate impacts are mainly driven by changes in the timing of Kettle River stages. In particular, peak flow in the river is expected to shift to an earlier date, and there will be a slightly prolonged and lower baseflow period. Portions of the valley aquifer that are strongly connected to the river will have the largest climate-driven changes. As the river peak flow shifts to an earlier date in the year, groundwater levels shift by the same interval. When comparing the difference in groundwater levels at the same day-of-year between 2040-2069 and current conditions, groundwater levels increase by up to 0.50 m near the river and up to 0.20 m away from the river. Increases in recharge under future climate scenarios are expected to be of minor in importance at this site.

The ability of a groundwater flow model to predict changes to groundwater levels, as forced by climate change, depends on the locations and types of model boundary conditions, the success of model calibration, and model scale. MODFLOW has limitations for modelling very complex aquifers, especially where there are perched water tables or where changes in the groundwater regime might be anticipated to cause changes to the surface water regime. The comparison of these two aquifers has demonstrated that different site-specific linkages exist for climatic impacts on groundwater resources. These can be successfully evaluated using standardized and consistent methodologies that allow for comparison of results and quantification of changes to groundwater levels, as well as for explanation for the causes of such changes.

6 CONCLUSIONS AND RECOMMENDATIONS

In this thesis I used two case studies of unconfined surficial aquifers to develop and apply a methodology for modelling aquifer recharge from downscaled GCM predictions. The modelled recharge for present and future climate scenarios was input to groundwater flow models to attempt to predict the climate change impacts on groundwater resources in these surficial aquifers, which were suspected of potential fast responses to such climatic changes. These two case studies shared common methodologies throughout and this facilitated a comparison of the model results at all steps of the modelling process. The common methodology for downscaling climate model results to local conditions, and then using weather generation to drive the recharge model, created a defensible and standardized methodology for generating recharge predictions for groundwater modelling projects.

Downscaling from Climate Models:

The main uncertainty still lies in the downscaling method performance, as demonstrated with large calibration bias between the downscaled present climate and observed present climate at a particular location. In effect, we do not know the actual future climate at any of the study location, but we are estimating it with imperfectly calibrated downscaling models, from also uncertain results of CGCM1 climate model.

The downscaling methods, that converted the outputs from the CGCM1 climate model, were not able to accurately simulate precipitation in the summer months at the interior mountainous location of Grand Forks aquifer.

Two different downscaling methods were used and compared, the SDSM model and the K-nn model. Both were statistical models that link the climatic variables in CGCM1 climate model to observed precipitation and temperature at a specific location (the two local climatic variables). At Grand Forks site neither SDSM nor K-nn adequately models precipitation, and the two models differed in goodness of fit to observed precipitation in different months of a year. One of the problems is that precipitation is not represented directly in a GCM, only humidity and other meteorological variables of the atmosphere being modelled. There are many local controls on precipitation (elevation, rain shadow effects, distance from ocean coast, etc.). The SDSM model results were selected because of better fit in the spring season to observed precipitation, when much of the recharge to surficial aquifer is thought to occur. However, at the coastal location of the Abbotsford-Sumas aquifer, the two downscaling models were much more consistent and had much smaller calibration bias to observed precipitation, and consequently, much greater confidence in precipitation predictions in future climate scenarios. At the interior location, summer precipitation was predicted to increase slightly, whereas at the coastal location summer precipitation was predicted to decrease slightly. The changes were not uniform at monthly time scales, but had seasonal differences.

In contrast to precipitation, the air temperature variable is directly represented in a GCM, including CGCM1 output. The predicted changes in monthly mean precipitation, and the associated changes in wet and dry spells, were realistically modelled at daily time scales using the LARS-WG stochastic weather generator, and then applied as inputs to the HELP infiltration (recharge to aquifer) model. The resulting downscaling model bias to observed mean daily air temperature was much less than for precipitation, in both SDSM and K-nn downscaling models. The predicted mean daily temperature had an increasing trend in all months from present to future scenarios by $1^{o} - 2^{o}C$ per 30 years, which reflects the CGCM1 results for that geographic region (CGCM1 grid square).

Recharge Modelling:

At the Grand Forks site, recharge results suggest that precipitation is insufficient to recharge the aquifer where there are thick sand and gravel terraces (i.e., on the elevated benches). Seasonal recharge has a bi-modal distribution in many recharge zones, and overall the model predicts that for the 2040-2069 scenario there will be 50% more recharge to the unconfined aquifer during the spring and summer seasons, compared to present climate scenario 1970-1999. In the autumn season, recharge is predicted to increase (10 to 25% depending on month within the season) or remain the same as present depending on location within the valley. In the winter, the CGCM1 predictions suggest less precipitation, and consequently, less recharge to aquifer. The predicted increase in recharge to aquifer will result, on average, in 0.2 m increase in groundwater elevation, although effects on future water supplies will be minimal.

The spatial distribution of recharge has consistently greater control on groundwater levels and flow rates than the temporal distribution of recharge in a groundwater flow model of the Grand Forks aquifer (dry climate), but it may be more significant in a wetter climate. If a mean annual recharge value is applied to the model, then modelled water levels are within 0.10 m of those calculated with temporally variable recharge applied at monthly intervals. The HELP model proved sensitive to several properties of the vadose zone; therefore, in order to achieve accurate results for recharge, the spatial variability of these key variables was considered in the development of recharge zonation maps for each study site. The improved resolution of recharge ensured that spatial distributions were accounted for in the analysis of climate change impacts, although the spatial distribution was of minor importance in the end at this particular aquifer.

At the Abbotsford-Sumas site, areal recharge is predicted to decrease by 5.6 to 6.3% relative to historic values under climate change for the 2010-2039 scenario. Greater decreases in recharge were predicted for the 2040-2069 climate scenario. The groundwater flow model results showed spatially-variable reduction in water levels ranging from 0.05 m to more than 0.25 m in most upland areas. In the 2040-2069 scenario groundwater level declines were also on the order of 0.25 m in most upland areas. These lower water levels will result in a reduction in hydraulic gradients from recharge to discharge areas, and a consequent scaled reduction in groundwater discharge. The lowering of the water table in the uplands area will most likely decrease baseflow in the streams fed mostly by seepage of groundwater. Lowland areas cannot be assessed because the model was constrained by specified head boundary conditions associated with major streams.

Groundwater - Surface Water Interactions:

With changing climate, the changes in surface water streamflows may locally affect groundwater flows in surficial aquifers. The GF aquifer is one example of such situation. Groundwater levels in the Grand Forks aquifer respond more directly to changes in the timing of basin-scale snowmelt events in the Kettle River, and the subsequent shift in the timing of the hydrograph, under scenarios of climate change, rather than to the changes in recharge. Between 11 and 20% of the river flows from the river into the aquifer during spring freshet, and storage duration in the alluvial aquifer lasts 30 to 60 days. Hydrologic modelling under scenarios of climate change suggests that Kettle River peak flow is expected to occur at an earlier date in the year and the baseflow period is expected to be of longer duration and lower than at present. The hydrograph shift for the 2040-2069 climate scenario is larger than in the 2010-2039 climate scenario, resulting in an apparent decrease in groundwater levels by up to 0.5 m during the spring season. In areas furthest away from the river influence, the direct precipitation recharge begins to dominate the response to climate change.

Groundwater modelling and monitoring should be continued and further scenarios evaluated that make predictions on changes in water consumption and climate together. The groundwater resources in the valley will not be affected significantly by these changes as long as the Kettle River maintains its discharge and supplies large quantities of recharge to the aquifer. Except near pumping wells, the aquifer groundwater levels cannot drop below the Kettle River water levels in the valley, even if there was no direct recharge from precipitation to the aquifer. In the end, the future groundwater use in the valley is limited by the withdrawal of an acceptable percentage of Kettle River discharge, especially at its minimum discharge rate in the late summer.

At the Abbotsford-Sumas aquifer further detailed investigations are required to measure the interaction of surface water and groundwater, through streamflow measurements and water level surveys, direct measurements of infiltration rates for different soil types to validate our recharge estimates, and better coupling of surface and groundwater in the flow models. Improvements to the model should consider changes in hydrology as a consequence to climate change, but more site-specific information on the streams and refinement of the model in those areas is needed. More detailed flow models will require much better surveying of stream channels and surveying of static water elevations in wells for the purpose of model calibration. Collection of soil permeability and infiltration rate data from many points in the valley should be collected to verify the recharge rates modeled in HELP.

Aquifer Heterogeneity:

The improved resolution and formulation of the model allowed better links with the river discharge/stage records and allowed for better representation of the river as specified head in the groundwater flow model (at greater spatial resolution). In areas with strong river-aquifer interactions, it is very important to adequately represent the aquifer heterogeneity in a groundwater flow model to accurately predict changes in groundwater levels. In the Grand Forks aquifer, we have a good idea what the effect of aquifer heterogeneity is on groundwater flow and climate impacts predictions but we don't know the actual aquifer heterogeneity, due to lack of high quality, high spatial density, and representative data for hydraulic conductivities.

The implications of not knowing enough about aquifer heterogeneities, where surface water interactions are dominant, is a large loss of model accuracy. The method of conceptual representation of aquifer heterogeneity in the model (influencing connectivity to the river) has uncertainty. This uncertainty is propagated through the modelling process and has as much influence on resulting modelled water levels (0.5 m changes in head) as do the predicted impacts of climate change on water levels. Therefore, in order to reduce these uncertainties, the spatial resolution of the model must be increased, thereby necessitating better resolution of aquifer heterogeneity.

Aquifer heterogeneity was also important in the characterization and modelling of the Abbotsford-Sumas aquifer (second study site). Due to the heterogeneity of the aquifer, the nature of interaction between the aquifer and the numerous streams could not be determined. The model could be re-calibrated if the representation of heterogeneities was improved. The locations of perched water tables should be investigated and the calibration data set modified. However, there are cost limitations and diminishing returns from collecting more data on hydraulic conductivity in many areas of the regional aquifer. Specific areas of interest should be identified and new data collected, in particular, the uplands near Abbotsford where groundwater flow model calibration was poor and where perched water tables are suspected to be present, and also along important streams draining south from the uplands west of Abbotsford. An improved understanding of groundwater chemistry and perhaps use of tracers to delineate capture zones would help to validate flow model results.

Final Comments:

As a result of all the uncertainties involved with climate downscaling, aquifer properties distributions, and model calibration, at this time the results of this study should be treated as a sensitivity study rather than actual predictions, even though we attempted to use "actual" best scientific guesses (model results) at future climate predictions, linked to the groundwater model through documented and defensible methodology presented in this thesis. This work showed one method of linking the groundwater flow models to climate model outputs and demonstrated that such links are practical to use in groundwater modelling studies, but that there are many uncertainties involved at each step. Climate model results downscaling to one location has large uncertainty and should be improved in future studies. The spatial distribution of recharge may be important in aquifers where recharge from precipitation dominates all other recharge pathways. Aquifer heterogeneity representation is always important and there is very large uncertainty involved when dealing with transient model behaviour, especially when river-aquifer interactions are important.

APPENDICES

Note: All appendices are included in electronic format (pdf documents) on attached CD.

Appendix 1: Recharge Modelling Methodology

Scibek, J., Allen, D.M., (2003 b). Groundwater Sensitivity to Climate Change (Part II): Analysis of Recharge for the Grand Forks Aquifer, Southern British Columbia. Report prepared for BC Ministry of Water, Land and Air Protection, 153 pp.

Appendix 2: Climate Change Impacts for Grand Forks Aquifer

Scibek, J., Allen, D.M., Whitfield, P., (2004). Groundwater Sensitivity to Climate Change (Part III): Climate Change Modelling Results for the Grand Forks Aquifer, Southern British Columbia. Report prepared for BC Ministry of Water, Land and Air Protection, 243 pp.

Allen D.M., Scibek, J., Whitfield, P., Wei, M., (2004). Climate Change and Groundwater: A Modelling Approach for Identifying Impacts and Resource Sustainability in the Central Interior of British Columbia. Report prepared for Natural Resources Canada, Climate Change Action Fund, 404 pp

Appendix 3: River-Aquifer Interactions at Grand Forks Aquifer

Scibek, J., Allen, D.M., (2003 a). Groundwater Sensitivity to Climate Change (Part I): Analysis of Watershed Water Balance and River-Aquifer Interactions for the Grand Forks Aquifer, Southern British Columbia. Report prepared for BC Ministry of Water, Land and Air Protection, 284 pp.

Appendix 4: Climate Change Impacts for Abbotsford-Sumas Aquifer

Scibek, J., Allen, D.M., (2005 a). Numerical groundwater flow model of the Abbotsford-Sumas aquifer, central Fraser Lowland of BC, Canada, and Washington State, US. Final report to Environment Canada, Vancouver, BC, 211 pp.

Scibek, J., Allen, D.M., (2005 b). Modelled Climate Change Impacts in the Abbotsford-Sumas Aquifer, Central Fraser Lowland of BC, Canada and Washington State, US. Proceedings of the Puget Sound-Georgia Basin Conference. March 30-April 1, 2005, Seattle, WA.

Scibek, J., Allen, D.M., (in prep). Groundwater Sensitivity to Climate Change: Abbotsford-Sumas Aquifer in British Columbia, Canada and Washington State, US. Prepared for: BC Ministry of Water, Land and Air Protection, and Environment Canada. Vancouver, BC, ~130 pp.

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