MODELING LOW-FLOW BEDROCK SPRINGS PROVIDING ECOLOGICAL HABITATS WITH CLIMATE CHANGE SCENARIOS

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Abstract

Groundwater discharge areas, including low-flow bedrock aquifer springs, are ecologically important and can be impacted by climate change. The development of and results from a groundwater modeling study simulating fractured bedrock spring flow are presented. This was conducted to produce hydrological data for an ecohydrological study of an endangered species, Allegheny Mountain Dusky Salamanders (*Desmognathus ochrophaeus*), in southern Quebec, Canada. The groundwater modeling approach in terms of scale and complexity was strongly driven by the need to produce hydrological data for the related ecohydrological modeling. Flows at four springs at different elevations were simulated for recent past conditions (2006-2010) and for reference (1971-2000) and future (2041-2070) periods using precipitation and temperature data from ten climate scenarios. Statistical analyses of spring flow parameters including activity periods and duration of flow were conducted. Flow rates for the four simulated springs, located at different elevations, are predicted to increase between 2% and 46% and will be active (flowing) 1% to 2% longer in the future. A significant change (predominantly an increase) looking at the seasonality of the number of active days occurs in the winter (2% to 4.9%) and spring seasons (-0.6% to 6.5%). Greatest flow rates were produced from springs at elevations where sub-horizontal fractures intersect the ground surface. These results suggest an intensification of the spring activity at the study site in context of climate change by 2050, which provides a positive habitat outlook for the endangered salamanders residing in the springs for the future.

Keywords: springs; bedrock aquifer; salamanders; climate change; discrete fracture network modeling; HydroGeoSphere

1. Introduction

Springs, a nexus between groundwater and surface water, play a vital role in the hydrologic cycle and provide critical ecological habitats. As expressions of subsurface flow, they maintain and support abundant ecosystems both near their outlets and downstream (Roy et al., 2011; Worthington et al., 2008; Barquín and Scarsbrook, 2008; van der Kamp, 1995), and they also provide important sources of surface water flow (Meyer et al., 2007; Boulton and Hancock, 2006; Smith et al., 2003; Merz et al., 2001). Considerable research has been conducted for many years to characterize flow mechanisms for high-flow springs (e.g., 1st, 2nd or 3rd magnitude flowing greater than 0.028 m^3/s) for anthropogenic uses including potable water supply and thermal baths (e.g., Malvicini et al., 2005; Bargar, 1978; Meizner, 1927), especially in karst environments (e.g., Brassington, 2007; Padilla et al., 2005; Fairleitner et al., 2005).

Low-flow intermittent (e.g., seasonal) or continuously flowing springs located in headwaters typically do not produce enough water for anthropogenic supply. For this reason, they are not a frequent research topic. However, these small springs provide important habitats for many plant and animal species (e.g., Wood et al., 2005). Knowledge about local seepage, or discharge, processes is also necessary to understand the dynamics of headwater streams which are the source of rivers and contribute to biodiversity (Meyer et al., 2007; Winter, 2007).

In the context of hydrological stressors such as climate change and increasing land development, it is critical to understand mechanisms that control spring flow to determine future viability of these important hydrological habitats. Regional or local-scale predictive studies about the impacts of climate change on groundwater resources are increasingly conducted. They show, for example, the future trends in hydraulic head and baseflow (or discharge) values stemming from a variety of recharge scenarios (e.g., Scibek et al., 2007; Jyrkama and Sykes, 2007). Climate change-induced variations in precipitation patterns (including volume and intensity) and evapotranspiration rates due to hotter temperatures, for example, can lead to shorter durations of spring activity and changing spring flow rates (e.g., Frisbee et al., 2013; Tambe et al., 2012; Rice, 2007). Seasonal changes in groundwater discharge are more pronounced for springs in headwater environments (where the flow paths are shorter) in contrast to discharge observed at the outlets of regional flow systems, which are more resilient to climate changes (Waibel et al., 2013; Dragoni and Sukhija, 2008).

Groundwater discharge, in the form of springs, can be represented using numerical models. These subsurface-focused or integrated (surface and subsurface flow) models can be used to conduct predictive scenarios to determine the effects of climate change, urbanization, or increasing groundwater use on small (and large) springs, which is required to implement mitigation measures to protect ecological habitats. Numerical modeling studies mostly focus on karstic settings in terms of interpretation of spring flow mechanisms using parsimonious lumped-parameter models (Hao et al., 2012; Amoruso et al., 2011; Barrett and Charbeneau, 1997; Bonacci and Bojanic, 1991) distributed and/or lumped parameter equivalent porous media models (Dragoni et al., 2013; Chen et al., 2013; Doummar et al., 2012; Scanlon et al., 2003); and channel flow (Eisenlohr et al.,

1997) or purely conduit flow (Halihan and Wicks, 1998) formulations. Equivalent porous media approaches for non-karstic fractured bedrock springs are also reported in the literature (e.g., Farlin et al., 2013; Swanson et al., 2006; Swanson and Bahr, 2004). In contrast to continuum and multi-continua approaches, discrete fracture network models in sedimentary and crystalline rock aquifer settings can be used to investigate the effects of individual fracture features on groundwater flow incorporating parameters including aperture, spacing, density and length (Voeckler and Allen, 2012; Levison and Novakowski, 2012; Blessent et al., 2011; Gleeson et al., 2009; Berkowitz, 2002).

Simulating small scale, low-flow springs using groundwater flow models in fractured bedrock aquifer settings that provide ecological habitats is challenging because the typical groundwater model discretization scale may be too large to represent individual springs. Moreover, fracture-dominated preferential flow introduces complexity that may not be adequately represented using equivalent porous media models. Groundwater flow modeling is often conducted at large scales (i.e., tens to thousands of meters per cell) to quantify for example hydraulic heads, river baseflows, groundwater renewal rates and residence times, or to understand anthropogenic impacts on groundwater resources (Levison et al., 2013; Sun et al., 2012; Michael and Voss, 2009). In ecohydrological modeling where connections to sensitive water-dependent habitats are made, it is important to accurately represent small-scale groundwater discharge features which may otherwise be overlooked.

This modeling study was driven by the need to obtain spring flow data (e.g., periods of activity) to be used as input to a salamander population model for an ecohydrological study (Larocque et al., 2013; Girard et al., 2014). The aim of this research is to simulate a typical headwater hillslope using a robust numerical model to better understand: 1) the hydrodynamics of small, low-flow bedrock aquifer headwater springs that support the habitat of endangered salamanders, and 2) the impact of climate changes on the spring dynamics. A fully integrated, or fully coupled, groundwater flow model is developed using HydroGeoSphere software (Therrien et al., 2010), a 3D model with discrete fracture and surface to subsurface simulation capability. The advantage of using an integrated model is that there is no need to estimate recharge separately (i.e., precipitation is divided into runoff and infiltration) (Brunner and Simmons, 2012). The modeling domain is based on a slice of a typical headwater catchment hillslope with numerous low-flow springs discharging from discrete fractures of a bedrock aquifer. This hydrological modeling representation and approach can be extended to other locations with similar topography and geology, and is especially important for investigating the sustainability of groundwater-dependent ecosystems. The results of this hydrological modeling have been applied to ecological modeling of salamander populations (Girard et al., 2014).

2. Materials and methods

2.1 Study area

The headwater hillslope used for this study is located in the Covey Hill Natural Laboratory on mount Covey Hill near the Canada-USA border in the Chateauguay River watershed (Larocque et al., 2006; Fig. 1). This field site has been used for several previous hydrogeological and ecological investigations (e.g., Lavoie et al., 2013; Levison et al., 2013; Gagné, 2010; Pellerin et al., 2009; Fournier, 2008). Covey Hill is the most northward extension of the Adirondack Mountains. It is a 20 km by 10 km E-W morphological feature (Nastev et al., 2008) and the highest elevation is approximately 345 m above sea level. The hill is mostly forested with limited areas of agriculture, including apple orchards and grazing.

Covey Hill comprises Cambrian sandstone of the Potsdam Group (Covey Hill Formation), deformed and fractured during the Appalachian orogeny (Globensky, 1986). The beds are relatively flat with horizontal to sub-horizontal bedding planes having dips of 1 to 5^{o} (Clark, 1966). The last ice advance (12 ky) eroded the surface deposits near the hilltop and south of the border. Locally the thin, permeable and sandy Saint-Jacques till is found on the hill (Lasalle, 1981). Glaciolacustrine sediments are found below 220 m above sea level (masl) and sandy beach deposits are located between 80 and 100 masl at the foot of the hill (Tremblay et al., 2010). Near the end of the last glaciation, the breakout of paleo lake Iroquois through an outlet near Covey Hill created a sandstone pavement (also called *Flat Rock*) that extends approximately 30 km southeastward into the Champlain Valley (Franzi et al., 2002). Covey Hill is considered an important recharge area for the Chateauguay aquifer (Croteau et al., 2010).

2.2 Hydrogeological conceptual model

A hydrogeological conceptual model of the Covey Hill Formation was developed by Nastev et al. (2008). This work forms the basis of the development of the discrete fracture numerical model used in the present study. The shallow bedrock aquifer is generally unconfined over the Covey Hill Natural Laboratory. Groundwater flows radially, generally to the north, from the hilltop predominately through bedding planes and joints. Flow through the very low permeability rock matrix is considered negligible. Following an extensive series of well pumping, packer and flow meter tests and an analysis of structural geology, Nastev et al. (2008) concluded that a permeable subhorizontal fracture is found every few tens to hundreds of meters. The lateral fracture continuity is hundreds of meters up to kilometers (Nastev et al., 2008). Similarly, Williams et al. (2010) found extensive lateral continuity in the sub-horizontal flow zones which are connected by high angle fractures in the Potsdam sandstones south of the Quebec-New York border. They commonly encountered horizontal flow zone spacing of less than 10 m. Groundwater discharge, in the form of low flowing springs, occurs where bedrock fractures intersect the ground surface (Nastev et al., 2008; Williams et al., 2010).

Contributing to the Nastev et al. (2008) study, Lavigne (2006) conducted straddle packer constant head injection tests (3.75 m intervals) in three monitoring wells drilled (40 to 76 m depth) into the Covey Hill Formation within the Covey Hill Natural Laboratory. Measured transmissivity (T) values within the packer zones range from 1.6×10^{-9} to 2.5×10^{-4} m^2/s. The geometric mean T is 9.3×10^{-7} $m^2/s \pm 1.6 \times 10^{-5}$ m^2/s. Using the Cubic Law (Snow, 1965) the minimum and maximum T correspond to single equivalent hydraulic fracture apertures of approximately 13 μm to 672 μm. On a more regional scale south of the Quebec-NY border, Williams et al. (2010) report T values in flow zones of the Postdam sandstones typically less than 1.1×10^{-4} m^2/s and up to 1.1×10^{-3} m^2/s.

The springs on Covey Hill provide habitat for the endangered Allegheny Mountain Dusky Salamanders (*Desmognathus ochrophaeus*) which are described further in Girard et al. (2013) and Larocque et al. (2013). Five bedrock springs on the northeast face of Covey Hill (Fig. 1) within the Natural Laboratory have been instrumented with water temperature loggers (Hobo) since 2007. These springs occur at elevations ranging between 140 to 180 masl. Thermographs can be used to interpret spring flow (Luhmann et al., 2011). Low temperatures and nearly sinusoidal variation in time indicates groundwater discharge while temperatures equal to the air temperature are interpreted as dry (i.e., a non-active spring) using the loggers. Fig. 2 illustrates a sample of a temporal series of spring water temperature and derived activity periods. This figure shows one spring which flows throughout the year (with attenuated measured temperatures) and an intermittent spring which flows generally from autumn to early summer and is dry during the hot summer months. The large temperature fluctuations generally observed in July and August, for example, are associated with the air temperature and not that of the groundwater. The activity periods of these two springs are typical of the monitored springs on Covey Hill. The continuously flowing spring is found at a lower elevation, where groundwater levels are relatively stable throughout the year, and the intermittent spring is located higher on the hill where groundwater levels show more temporal variations. Annual variations in groundwater levels have been shown to increase with elevation on Covey Hill (Levison et al., 2013). It is expected that spring intermittency also increases with elevation.

Measuring flow rates in small springs can be extremely difficult. Due to the low-flow nature of the springs, very few flows were measured on Covey Hill, with the exception of a rate for one spring measured at 9.0×10^{-5} m^3/s in May 2011 using a stopwatch and a graduated container. It is assumed that the flows at the time of snow melt (i.e., earlier in the spring season) are higher than this. The flow rates of the springs are assumed to be one or more orders of magnitude lower than the baseflow of the smallest gauged stream in the area (Schulman stream, Fig. 1) which has been estimated to vary from 1.0×10^{-3} m^3/s to 2.4×10^{-2} m^3/s (Levison et al., 2013). On the northeastern face of Covey Hill the spring outflow zones are generally 15 m or greater in length and 0.5 to 10 m in width (Bilodeau, 2002).

2.3 Weather data and climate change scenarios

Precipitation and temperature data are available from the Hemmingford weather station located approximately 11 km from the study area (Environment Canada, 2010). Snow usually falls between November and March. From 1971 to 2000 the average annual temperature was 6.4°C. Over this time period the average monthly temperature ranged from -9.6°C in January to 20.6°C in July with a winter minimum of -30.5°C and a summer maximum of 30.3°C. The average monthly and total rainfall was 73 mm and 872 mm, respectively (Environment Canada, 2010).

The hydrogeological model constructed for this study requires the use of net precipitation (Pnet, which is precipitation minus evapotranspiration) as an atmospheric water input. The model directs Pnet to surface flow (runoff) and infiltration (recharge). Levison et al.

(2013) used potential evapotranspiration (ET) calculated using the Oudin et al. (2005) equation. This gives ET estimates based on mean daily air temperature and on extraterrestrial radiation, estimated following Morton (1983). The same approach was used in the present study. The Pnet is estimated for use in the numerical hydrogeological model as described in the following section. Pnets were calculated on a monthly basis. This permitted the identification of the months during which Pnet is often zero due to the non-availability of water for evapotranspiration. A negative net precipitation indicates a month where potential evapotranspiration could not be met by precipitation, such as some summer months. Because of the sub-zero temperatures, the winter net precipitation accumulates on the ground as snow and becomes available during snowmelt in the spring season. For comparison purposes, the Pnet values in Levison et al. (2013) were calculated on an annual basis, instead of monthly. In that work, the annual calculation lead to Pnet changes from the reference period (1971 to 2000) to the future period (2041 to 2070) ranging from a 30% decrease to a 10% increase. The bulk annual calculation incorporates the negative Pnet values encountered during the summer. For the monthly calculations herein, periods of restricted water availability (i.e., summer months when evapotranspiration is greater than precipitation) the Pnet values were set to 0 for use in the numerical model. Pnet values calculated herein are therefore higher than those reported in Levison et al. (2013). The average annual Pnet from 1971 to 2000 using data from the Hemmingford weather station was 408 mm.

The climate change scenarios from Levison et al. (2013) were used for this study. Similarly to the work of Levison et al. (2013), the impact of climate change on the system was investigated by calculating Pnet values with future time series of daily precipitation and temperature data. Predicted ET values were derived from the most recently available Regional Climate Models (RCMs), using future temperature time series in the Oudin et al. (2005) equation. The climate change scenarios are derived from four RCMs driven by six General Circulation Models (GCMs). Future RCM scenarios were further downscaled using the daily translation bias correction method (Mpelasoka and Chiew, 2009) to remove the biases between simulated and observed temperature and precipitation variables.

Ten projections were selected from the 25 dynamically downscaled simulations available for the Covey Hill area (see Table 1 for annual synthesis and Fig. 3 for monthly variations). Most of the simulations are outputs of the Canadian Regional Climate Model (CRCM) (Music and Caya, 2007) and were generated and supplied by the Ouranos Consortium on Regional Climatology and Adaptation to Climate Change. The remaining simulations are from the North American Regional Climate Change Assessment Program. All projections are for the 2050 horizon (2041-2070). The 10 simulations account for 85% of the future climate variability projected for the study site as established by a cluster analysis carried out on the range of available RCM scenarios. The simulations are driven by six different GCMs under the Intergovernmental Panel on Climate Change emissions scenarios A1B and A2 (IPCC, 2000).

The outputs of climate models, once corrected for their bias, satisfactorily reproduce the average monthly temperature and precipitation for the reference period (1971-2000). The difference between the observed annual average temperature and that simulated by the climate models over the 30 years is 0.4°C. The difference between observed and simulated average annual rainfall was 5.5% (Larocque et al., 2013). The ensemble mean of future (2041-2070) simulations predicts an increase of 1.8 (March) to 3.0°C (January) for monthly temperatures (Fig. 3a). This increase depends on the particular month, but the envelope of uncertainty remains relatively equal throughout the year. The ensemble mean of future simulations predicts an increase in monthly precipitation for every month except June (Fig. 3b). Although most of the climate models predict an increase in precipitation during the winter, the envelope of uncertainty remains high (-3 to 47%), and the signal is mixed for the summer and fall months.

Stemming from these changes in temperature and precipitation, the ensemble mean of calculated Pnet values shows a maximum increase of 45 mm in March and a maximum decrease of -48 mm in April. It is important to note that the envelope of uncertainty during these two months is very high: -5 to 109 mm in March and -94 to 14 mm in April. These spring season-related maximum variations can be linked to higher winter temperatures and earlier snowmelt. An earlier onset of spring and more winter recharge have been reported in other studies as effects to be expected from climate change (e.g., Waibel et al., 2013). Pnet variations are small during the summer period when available precipitation (for runoff and infiltration) is usually very low due to high temperature and significant evapotranspiration. The width of the envelope is minimal from May to September, and close to zero from May to June, suggesting that the future Pnet will not be very different from the reference period.

2.4 Development of the hydrogeological model

A three dimensional fully integrated discrete fracture numerical model was constructed using HydroGeoSphere (Therrien et al., 2010; Brunner and Simmons, 2012) for the purpose of simulating spring flow at various elevations for the reference period and future predicted climate scenarios. The modeling approach in terms of scale and complexity was strongly driven by the need to produce hydrological data to simulate the local dynamics of salamander habitats. The Allegheny Mountain Dusky Salamanders have a m-scale home range size (Holomuzki, 1982), which means they live and travel within a very small (e.g., 1 m^2) area

For the entire Covey Hill Natural Laboratory a finite difference model was first developed by using MODFLOW to represent interactions between a headwater peatland and the surrounding bedrock aquifer (Levison et al., 2013). For the present study, it was determined that a smaller cell size, closer to that of the salamander home range of 1 m x 1 m, was needed to represent the small bedrock springs that are salamander habitats. The MODFLOW model used cells of 135 by 135 m and covered a large area of 173 km^2. The flexibility of changing the discretization and the need to represent the bedrock springs more realistically from a physical perspective using discrete fractures led to the development of the local scale model using HydroGeoSphere. The discrete-fracture integrated model allows water to flow freely to the surface in distinct locations where fractures intersect the surface (i.e., the discharge is not constrained by the placement of drain nodes or seepage faces). During the model formulation, it was found that to get distinct small discharge areas akin to those observed in the field as bedrock springs, an impermeable rock matrix with discrete fractures that intersect the ground surface were required. The fractures that intersect the surface perform both recharge and discharge functions: 1) precipitation can infiltrate through the fractures into the aquifer; and 2) groundwater can discharge through the fractures as spring flow.

This transient model developed using HydroGeoSphere simulates surface runoff as well as the flow in the unsaturated and saturated fracture network, across a range of Covey Hill elevations where bedrock springs, providing salamander habitats, are known and instrumented along the NE face of Covey Hill (Fig. 1). The model domain is 4500 m along the x-axis (approximately SW to NE), 100 m along the y-axis (roughly SE to NW), to a depth of 100 m below the ground surface on the z-axis. Elevations vary only along the x-axis and not along the y-axis (see Fig. 4). Surface elevation changes from 330 m above sea level to 85 m above sea level to the NE. The rock matrix is considered impermeable, which means that precipitation recharges to and discharges from the fractures in the bedrock aquifer. The southern (highest elevation) and lateral boundaries of the model are set as zero flux since the southern boundary is at a topographical divide and the model is considered to be oriented generally along a flow line. The northern boundary (lowest elevation) of the model is set as constant head equal to the regional piezometric surface at that location.

The previously described conceptual model (Nastev et al., 2008) including the detailed packer tests (Lavigne, 2006; Lavigne et al., 2010) were used to develop the numerical model. During the model construction and calibration phase, the fracture apertures and spacing were modified within the field measured and conceptualized range of values to meet the following general calibration objectives: 1) obtain discharge generally at locations where springs are observed on Covey Hill (i.e., in the area between x = 1800 to 2400 m in the model); 2) produce the dynamic range of both the intermittent and continuous characteristic spring flow, as observed in the field from temperature loggers monitoring spring activity (Fig. 2); and 3) achieve reasonable flow rates discharging to the surface from fractures (that is, less than measured baseflow in the smallest gauged stream in the Natural Laboratory). Fracture apertures were modified to achieve realistic recharge values (e.g., similar to those reported in Levison et al., 2013).

The model was run using the Covey Hill Natural Laboratory average annual Pnet of 372 mm from the past decade (2000-2010) until a steady-state was achieved. This provided a means to determine the ballpark aperture sizes and spacing that yielded reasonable infiltration rates during a period for which field-measured variables (activity period of springs and Schulman stream flow rates) were available. Values suggested in the HydroGeoSphere manual (Therrien et al., 2010) were used for the unmeasured surface flow parameters. The parameters used in the calibrated model are summarized in Table 2. This model was constructed to generally represent the topography and conceptual model of the fractured bedrock aquifer, while maintaining a simple and not overly-constrained model formulation.

Following the model calibration in steady-state, transient simulations were performed first using the monthly Pnet calculated from the measured Hemmingford weather station data for the recent past (2006-2010), for the 1971-2000 period and for the 10 climate change scenarios. These monthly Pnet values were applied directly to the surface of the model using a specified flux boundary condition to represent the net precipitation reaching the ground surface. The model then divides the surface runoff and infiltration based on the properties of materials and their saturation. In transient-state the model ran with a variable hydraulic head-dependent time step, producing flow result output more frequently than once per day. Following the 1971-2000 run, transient state simulations were carried out for the 10 reference and 10 future scenarios using monthly Pnet data from T and P values produced by the climate models.

Detailed flow out of the vertical fractures at x = 1800 m (z = 177 m), x = 2000 m (z = 162 m), x = 2200 m (z = 150 m), and x = 2400 m (z = 140 m) were simulated and analyzed. Springs located at z = 177 m and 140 m are located at or near to the intersection of sub-horizontal fractures with the surface topography.

3. Results

3.1 Recent past

The vertical and horizontal apertures are 600 μm and vertical fractures intersect the surface at a spacing of 200 m throughout the domain. This configuration produced the best balance between the calibration objectives stated previously, within the range of available field data (e.g., fracture apertures; observed spring locations). For calibration objectives 1 and 2, over the recent past (2006-2010), the calibrated model was able to produce the characteristic spring flow observed in the field (Fig. 5). That is, there are both continuously flowing and intermittently flowing springs at different elevations in the model. In the model, the springs at lower elevations flow for longer durations and the length of spring activity generally decreases with increasing elevation. The springs flowing in the model reside between z = 177 m and 140 m in elevation, where they are also observed in the field. For objective 3, the maximum spring flow for the 2006-2010 simulation was approximately 3.8×10^{-4} m^3/s, more than one order of magnitude less than the baseflow of the smallest gauged stream in the area (Schulman stream, Fig. 1), estimated to vary from 0.001 to 0.024 m^3/s (Levison et al., 2013). This is considered reasonable since seepage from the flow of the observed and instrumented individual bedrock springs is visually insignificant compared to the flow of the gauged Schulman stream during minimum flow (baseflow) conditions.

The calibrated steady-state recharge values from the regional MODFLOW model of Levison et al. (2013) over the location of the current model domain range from 0% of Pnet in the extreme NE (surficial deposits are clay) to a maximum of 88% of Pnet at the hilltop, with an average of 37% of Pnet. The calibrated current model, when run for a sufficient amount of time to reach a steady-state, had an infiltration value of 31.9% of the water applied to the model surface (i.e., the Pnet). Interestingly, the spring discharge (or model exfiltration), which was not simulated in the regional model of Levison et al. (2013), is equal to 15.6% of the Pnet. This leaves 16.3% of the total Pnet that infiltrates and flows through the deep horizontal fracture (Fig. 4) to the northern constant head boundary as groundwater recharge to the regional fractured bedrock aquifer. The regional recharge in complex headwater systems is typically difficult to determine. This integrated numerical modeling approach that simulates groundwater discharge allows for the quantification of regional recharge in this headwater area. The remaining 83.7% of the water is therefore surface water runoff and leaves the model through a critical depth boundary condition set on the lateral and northern boundaries. Table 3 shows example changes in vertical aperture sizes and their effect on the model performance. This is for steady state simulations using an average Pnet for the past decade (2000-2010) of 372 mm/year. The 600 μm run was chosen in contrast to the 500 μm run, for example, because the infiltration rate was closer to the target value coming from the regional Levison et al. (2013) calibrated recharge value of 37% of Pnet.

Fracture aperture (opening) effects both recharge and discharge processes. It has a large effect on the amount of water that can infiltrate and reach the deep subsurface. For example, Fig. 6 shows the percent of the applied Pnet that recharges the aquifer to reach the northern constant head boundary (at steady-state) for apertures ranging from 400 to 1000 μm, with all other properties being held equal. For infiltration (not necessarily deep recharge, since a portion of this water can discharge as springs) 500 to 700 μm aperture changes cause infiltration to increase (24.7% to 38.7% of the same applied Pnet) and exfiltration (i.e., spring discharge) to conversely decrease (15.2% to 12.9% of the Pnet) as shown in Table 3. In terms of the discharge rate, however, increasing the apertures causes a higher peak spring flow. For example, with 500 μm apertures the maximum flow using 2010 Pnet data was 2.4×10^{-4} m^3/s, while for 700 μm apertures it more than doubled to

5.6x10^{-4} m^3/s (Table 3). Holding the horizontal fractures the same size and changing vertical aperture has a similar effect. For example, with 700 µm horizontal fractures the maximum spring flow using 2010 Pnet data ranges from 4.6x10^{-4} m^3/s to 5.6x10^{-4} m^3/s for 500 to 700 µm vertical fractures, respectively (Table 3). Holding the vertical fractures the same size (800 µm) and increasing horizontal fractures from 600 to 800 µm induced a decrease in exfiltration percentagewise (total spring discharge) from 15.6 to 6.9% of the applied Pnet. For similar headwater systems with low-flow springs, larger apertures can be expected to promote aquifer recharge with potentially higher maximum spring flow rates, but having a lower total volume of spring discharge.

3.2 Climate change scenarios

The numerical model was also run from 1971-2000 using observed data and for 10 climate change scenarios (for both reference and future periods) using data produced from climate models. Statistical analysis of the spring flow magnitude, duration and seasonality was conducted for the thirty year periods. The variables studied are: 1) the average flow rate of the springs (when they are active); 2) the number of flowing (active) days per year; 3) the average duration of spring activity; and 4) the seasonal distribution of spring activity. Each variable is discussed in the following paragraphs. Table 4 summarizes the results of the statistical analyses in terms of the ensemble averages.

Fig. 7 shows the flow of springs at four different elevations (z = 140, 150, 162 and 177 m) simulated using Pnets: 1) from the observed meteorological data at the Hemmingford weather station (OBSERVED) for 1971-2000; and 2) calculated with the rainfall and temperature data of 10 climate scenarios for the period 1971-2000 (named REF) and for the period 2041-2070 (named FUT). The simulated flows for all climate scenarios are the same order of magnitude as those produced using the Hemmingford weather data for the 1971-2000 period. Generally, the springs at z = 177 m (the highest elevation with a flowing spring) and 140 m have rates one order of magnitude higher (2.0×10^{-4} and 2.6×10^{-4} m^3/s) than the springs at z = 162 and 150 m (mid-slope) where the average rate is 2.4×10^{-5} and6.1×10^{-5} m^3/s (ensemble means) for the future.

It could be expected that flow rate varies according to the elevation of the spring, with springs at lower elevations producing greater flows. However, in a porous medium, the flow rate is a function of the hydraulic conductivity, the hydraulic gradient, and aquifer thickness and width. Non-homogeneous hydraulic conductivity, varying aquifer thicknesses, or lower gradients may lead to smaller discharge at times in springs at lower elevations. However, during long periods without recharge higher elevation springs are generally more vulnerable to drying up. For fractured media specifically, the characteristics of the local fracture network must also be considered regarding spring flow rates and discharge locations (e.g., Di Matteo et al., 2013). An equivalent porous media model could reproduce the flow if the fracture is replaced by a volume of similar shape having high hydraulic conductivity and effective porosity. However, using a model that can simulate discrete fracture flow directly is more efficient and may be better suited to simulate the tarissement stage of the spring flow. This speaks to the importance of using discrete fracture numerical modeling for certain geological conditions and modeling applications, especially for ecohydrological investigations requiring detailed input data for population modeling. Thus, the greatest spring flows observed at 140 and 177 m above sea level can be explained by the fact that they are located where or near to locations where major sub-horizontal fractures intersect the surface, unlike two mid-slope springs coming from vertical fractures intersecting the surface (Fig. 4). The greatest flow is observed for the spring at the lowest elevation (Fig. 7a). The maximum flow rate is 4.3×10^{-4} m^3/s, while the minimum flow rate of 1.0×10^{-8} m^3/s is observed at the spring at z = 162 m. The spring flows are on the same order of magnitude as the field measurement and are at least one order of magnitude lower than the baseflow of the smallest gauged stream in the watershed. Flows rates for the four simulated springs are likely to increase in the future, which was simulated for all of the climate scenarios. This increase, which is significant using a Wilcoxon-Mann-Whitney test for paired samples ($\alpha = 0.05$), is of the order of 5 to 6% depending on the selected spring (Table 4).

For the number of days of spring activity, the observed values range from 10 for the spring at 162 m to 275 for the spring at 140 m, which are in the range of values for the simulated climate change scenario reference period. The activity of simulated springs varies considerably from one spring to another, with the lowest elevation spring being active on average more than 75% of the time and the spring at 162 m being active for only 3% of the year (Fig. 8). Like the spring flow magnitude, the presence of the sub-horizontal discharging fracture at 177 m elevation explains the greater activity than at z = 162 m. The average number of days of activity for the four springs (from lowest to highest elevation) for the reference period is 282, 57, 12 and 24 days. The results for the future period are 289, 66, 18 and 29 days, indicating an increase in the number of days of activity for all springs, although this increase varies considerably depending on the spring (e.g., 2% for the spring at z = 140 m and 46% for the spring at z = 162 m). The differences observed between the reference period and the future period are significant according to the Wilcoxon-Mann-Whitney test for paired samples ($\alpha = 0.05$). The spring at the lower elevation (140 m) is most active, while the least active spring is generally the one at 162 m. The maximum difference of days of activity is usually found between scenarios MRCC4.2.3_ECHAM#1 (wetter) and MRCC_CCSM (drier).

The results from the climate scenarios for the average duration of activity (length of consecutive flowing days) generally agrees well with those simulated from observed Hemmingford weather data from the 1971-2000 period. There is only one spring (at 150 m elevation) where the variability in the observed data exceeds that of climate scenarios (Fig. 9b). Similarly to the number of days of spring activity, the longest period of activity is observed at the lowest elevation spring, where a period of activity in the reference period is an average of 55 consecutive days (Fig. 9a; Table 4). The spring at 162 m elevation has the shortest activity periods. For the future climate, the signal is mixed. The two springs at higher elevations show an increase of about 2 to 3 days while the two springs at the lower elevations have a shorter period of activity by 1 to 2 days. With the exception of the spring located at 162 m, the differences between the reference period and the future period are not significant ($\alpha = 0.05$) using a Wilcoxon-Mann-Whitney test for paired samples. Considering the least optimistic case, the 144 m elevation spring is predicted to have a flow period length that is 10 days shorter, and the 177 m elevation spring is predicted to have a flow period length that is 6 days shorter (both for the MRCC4.2.3_CBCM3#5 projection).

Fig. 10 shows the seasonal distribution of spring flow. These results represent the average of the sum of days of activity per season over 30 years. For all springs, the differences are significant between the reference period and the future period for the winter season (DJF) and spring (MAM) using a Wilcoxon-Mann-Whitney test for paired samples ($\alpha = 0.05$). The differences observed for summer (JJA) and autumn (SON) seasons are contrastingly not significant. For each of the springs, the number of active (or flowing) days during the winter increases, a potential indication of earlier spring thaw. In the spring, the number of days of activity also increases, with the exception of spring at the highest elevation (Fig. 10d). For each of the springs, the number of flowing days during the summer and autumn are lower in the future compared to the reference period.

In summary, using the numerical model to represent spring flow for past conditions and for reference and future climate change scenarios, there was: 1) a significant increase in the average rate of flow for all springs; 2) a significant increase in the number of days of activity for all springs; 3) no significant impact of climate change on the length of periods of activity, with the exception of the spring at 162 m elevation; 4) a significant increase in the number of active (flowing) days in the winter and spring seasons for all springs, and 5) insignificant changes for the summer and fall seasons.

These results suggest an intensification of the spring activity on Covey Hill in context of climate change by 2050, which provides a positive habitat outlook for the endangered salamanders residing in the springs for the future. This increase is significant both in terms of quantities of spring flow and the number of days when flow occurs. This increased activity can be attributed to, among other processes such as increased Pnet, the shortening of the winter period and, therefore, the earlier arrival of spring melt.

4. Conclusions

This research focused on the simulation of flow rates in small low-flow bedrock aquifer headwater springs and on the impact of climate changes on the spring dynamics. An integrated numerical flow model was developed for a typical headwater hillslope. The groundwater model was used to simulate past (2006-2010 and 1971-2000) and future (2041-2070) spring flow, based on historical data and input produced by numerous Regional Climate Models.

During the flow model development, a discrete fracture representation and smaller-than-typical scale groundwater modeling were found to be useful to physically represent discharge at individual springs from a shallow fractured sandstone aquifer with a low permeability matrix and sub-horizontal bedding planes. Equivalent porous media numerical modeling approaches may not be suitable to capture the fineness and discrete spatial formulation required to be coupled with ecological habitat modeling studies for small organisms such as salamanders, which was the driving force behind this research. For groundwater flow modeling studies that are coupled with other models, either biological or from other scientific fields, it is critically important to frame the groundwater modeling approach to best fit the required inputs (e.g., scale in space and time and data formats) or desired outcomes of the coupled model. Thus, innovative formulations and balancing between domain size and fineness of representation must be a strong consideration during groundwater or hydrological model development for ecohydrological studies.

Greatest flow rates were produced from springs at elevations where sub-horizontal fractures intersect the ground surface. The model was able to produce, in one case, higher average flow rates from a high elevation spring compared to other two lower elevation springs. This can occur where sub-horizontal fractures intersect the ground surface. Certain physical subtleties afforded with discrete fracture network modeling cannot be represented using lumped equivalent porous media approaches. Thus, the importance of discrete fracture flow cannot be overlooked for certain cases, such as investigating individual groundwater discharge features. The small bedrock springs of Covey Hill were found to flow more, both in terms of magnitude and days of flow, in the future than in the past. For this location, and in similar geographical, climate and topographical contexts, hydrological features supporting habitats or important for water supply may have increased flow in the future, as predicted using the input data from the Regional Climate Models. However, other anthropogenic impacts such as increasing population and more water withdrawals could reverse this increasing trend.

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Table 1. Climate scenarios selected for this study, simulated air temperature and precipitation changes, and derived net precipitation changes.

RCM	Driven by this GCM	Member	Domain	Emission scenario	Air temperature 1971-2000 (°C)	Air temperature 2041-2070 (°C)	Air temperature change (°C)	Precipitation 1971-2000 (mm)	Precipitation 2041-2070 (mm)	Precipitation (% change)	Pnet 1971-2000 (mm)	Pnet 2041-2070 (mm)	Net precipitation (% change)
CRCM4.2.3	CGCM3	5	AMNO	A2	6.8	9.9	3.1	922	1000	8	418	473	13
CRCM4.2.3	CGCM3	2	AMNO	A2	6.7	9.9	3.2	922	1003	9	421	470	12
CRCM4.2.3	ECHAM5	1	AMNO	A2	6.8	9.0	2.2	922	1030	12	418	480	15
CRCM4.2.3	ECHAM5	2	AMNO	A2	6.7	9.3	2.5	922	1017	10	420	445	6
CRCM4.2.3	Arpège UnifS2	--	AMNO	A1B	6.7	8.7	1.9	922	989	7	420	456	8
CRCM4.2.0	CGCM3	4	AMNO	A2	6.7	9.5	2.8	922	977	6	417	436	5
CRCM	CCSM	--	N. Amer.	A2	6.8	9.7	3.0	920	934	2	429	410	-4
ECP2	GFDL	--	N. Amer.	A2	6.7	9.3	2.6	922	1031	12	426	472	11
HRM3	HADCM3	--	QC	A2	6.3	9.0	2.8	908	992	9	430	463	8
RCM3	CGCM3	--	N. Amer.	A2	6.7	9.4	2.7	922	953	3	426	425	0
Ensemble mean					6.7	9.4	2.7	920	992	8	422	453	7

Table 2. Model parameters.

General properties	Value
Model length	4500 m
Model width	100 m
Model thickness	100 m (highest elevation: 330 m; lowest elevation: 85 m)
Cell size	20 m x 20 m in *x* and *y* (variable in *z*)
Northern boundary condition (lowest elevation)	Specified hydraulic head at 75 m (10 m below the surface)
Southern boundary condition (highest elevation)	Zero flux
Eastern boundary condition	Zero flux
Western boundary condition	Zero flux
Top boundary condition	Variable flux (precipitation) – monthly Pnet values used
Fracture characteristics	**Value**
Vertical fracture spacing	200 m (along *x*)
Vertical fracture apertures	600 μm
Horizontal fracture apertures	600 μm
Rock matrix characteristics	**Value**
Hydraulic conductivity	$1\text{x}10^{-20}$ m/year (essentially impermeable)
Porosity	0.001
Surface flow characteristics	**Value**
x and *y* friction factors	$1.585\text{x}10^{-9}$
Rill storage height	0.001 m
Coupling length	$1\text{x}10^{-4}$ m

Table 3. Various fracture apertures and their effect on model performance (all with 200 m vertical fracture spacing). Applied Pnet was the average value (372 mm/year) for the recent past (2000-2010).

Horizontal aperture	Vertical aperture	Continuously* flowing spring at z = 140 m (2010)?	Max spring flow rate (2010) (m^3/s)	Infiltration (% of applied Pnet)	Exfiltration (% of applied Pnet)	Flow through constant head boundary (% of applied Pnet)
500	500	continuous	2.4E-4	24.7	15.2	9.4
700	700	mid-Feb:end June; end July:continuous	5.6E-4	38.7	12.9	25.9
700	500	mid-Feb:end June; mid Aug-end Nov	4.6E-4	30.3	10.4	19.9
800	600	end Feb:mid-March; May; Aug:mid-Sept	6.2E-4	37.6	6.9	30.7
700	600	mid-Feb:end June; Aug-mid-end Nov	5.1E-4	24.4	11.6	22.8
600**	600	continuous except for short period in February	3.8E-4	31.9	15.6	16.3

*** all configurations produced intermittently flowing springs**

**** calibrated model**

Table 4. Simulated spring flow results (ensemble averages) for the four elevations of interest. REF = reference period and FUT = future period.

		140 m		150 m		162 m		177 m	
		REF	FUT	REF	FUT	REF	FUT	REF	FUT
Average flow (10^{-4} m^3/s)		2.26	2.41*	0.57	0.61*	0.24	0.26*	1.81	1.93*
Number of active days of flow (d)		282	289*	57	66*	12	18*	24	29*
Duration of activity periods (d)		55	53	24	23	10	13*	14	16
Seasonal partitioning (%)	DJF	23.1	25.1*	13.8	17.5*	7.9	10.8*	7.2	12.1*
	MAM	25.4	27.7*	59.1	61.8*	76.9	83.4*	78.4	77.8*
	JJA	28.1	25.7*	11.8	7.8	3.2	0.3	2.3	0.4
	SON	23.5	21.4	15.2	12.9	12.0	5.5	12.2	9.7

* Significant difference (with α=0.05) using the Wilcoxon-Mann-Whitney test for paired samples.