Hydrological sensitivity of a northern mountain basin to climate change

Kabir Rasouli,1 John W. Pomeroy,1* J. Richard Janowicz,2 Sean K. Carey3 and Tyler J. Williams1

1 Centre for Hydrology, University of Saskatchewan, Saskatoon, SK, Canada
2 Yukon Environment Water Resources Branch, Whitehorse, YT, Canada
3 School of Geography and Earth Sciences, McMaster University, Hamilton, ON, Canada

Abstract:

The hydrological sensitivity of a northern Canadian mountain basin to change in temperature and precipitation was examined. A physically based hydrological model was created and included important snow and frozen soil infiltration processes. The model was discretized into hydrological response units in order to simulate snow accumulation and melt regimes and basin discharge. Model parameters were drawn from scientific studies in the basin except for calibration of routing and drainage. The model was able to simulate snow surveys and discharge measurements with very good accuracy. The forcing inputs of the hourly air temperatures and daily precipitation were scaled linearly to examine the model sensitivity to conditions included in a range of climate change scenarios: warming of up to 5 °C and change in precipitation of +/− 20%. The results show that peak seasonal snow accumulation, snow season length, evapotranspiration, runoff, peak runoff, and the timing of peak runoff have a pronounced sensitivity to both warming and precipitation change, where the impact of warming is partly compensated for by increased precipitation and dramatically enhanced by decreased precipitation. The snow regime, including peak snow accumulation, snow-free period, intercepted snow sublimation, and blowing snow transport, was most sensitive to temperature, and the impact of a warming of 5 °C could not be compensated for by a precipitation increase of 20%. However, basin discharge was more sensitive to precipitation, and the impact of warming could be compensated for by a slight increase in precipitation. The impacts of 5 °C warming with a +/− 20% change in precipitation resulted in snow accumulation, runoff, and peak streamflow decreasing by from one half to one fifth and the snow-free period lengthening by from 46 to 60 days; in both cases, the smaller change is associated with increased precipitation and the larger change with decreased precipitation. These results show that mountain hydrology in Northern Canada is extremely sensitive to warming, that snow regime is more sensitive to warming than streamflow and that changes in precipitation can partly modulate this response. Copyright © 2014 John Wiley & Sons, Ltd.

KEY WORDS climate change impact; Yukon; sensitivity analysis; snow hydrology; cold regions; hydrological modelling

Received 11 November 2013; Accepted 6 May 2014

INTRODUCTION

Air temperatures in northern latitudes in general and specifically in the Yukon Territory, Canada have increased significantly within the last decades, while changes in precipitation are more variable with summer precipitation generally increasing throughout Yukon, but winter precipitation decreasing in southern Yukon and increasing in central and northern regions (Janowicz, 2010). These temperature and precipitation changes are associated with changes in the hydrological cycle (Bunbury and Gajewski, 2012), such as the river ice cover period shortening due to later freeze-up in the fall, and earlier ice break-ups in the spring and a greater frequency of mid-winter break-up events with subsequent flooding (Janowicz, 2008, 2010). A warming climate can result in permafrost degradation, expansion of shrub tundra (Tape et al., 2006), and altered snow processes (Pomeroy et al., 2006), each of which impacts streamflow (St Jacques and Sauchyn, 2009). The changing shrub extent, which might enhance or delay melting of permafrost in tundra regions (Sturm et al., 2005) and its relation to the surface energy balance alters albedo feedbacks to the climate system (Pomeroy et al., 2006). Studies show that river flow in winter and April has also increased in Yukon and regions of significant permafrost with ground thaw (Walvoord and Striegl, 2007). In April, a transition period between streamflow dominated by baseflow and snowmelt runoff, groundwater contribution to streamflow has increased (Brabets and Walvoord, 2009). The changes in the ice/snow cover, permafrost, streamflow, soil moisture, vegetation, and other defining properties of ecosystems include important feedback processes transforming the ecology, surface energy balance, and hydrological cycle of northern Canada at local to global scales (Ostertamp et al., 2009; Rawlins et al., 2009).

In northern latitudes, alpine and shrub tundra snow is redistributed by wind; the blowing snow transport and
sublimation processes are affected by the interaction of local topography and landscape vegetation cover with regional wind flow patterns (Pomeroy et al., 1999; MacDonald et al., 2009). Snow is intercepted in needle-leaf forest canopies from which large proportions of winter snowfall sublimate rather than unload to the surface (Pomeroy et al., 1999). High surface runoff derives from spring snowmelt and occurs as a result of limited infiltration into frozen mineral soils at the time of melt and a relatively rapid release of water from melting snowpacks (Janowicz et al., 2003). Snowmelt timing and melt rate are primarily controlled by the net inputs of solar radiation, longwave radiation, energy advected from rainfall, and turbulent transfer of sensible and latent heat (Pomeroy et al., 2003). The impact of these inputs on snowmelt are moderated by the storage of internal energy in the initially cold snowpacks and the snow surface albedo, both of which change rapidly in the pre-melt and melt period. Snowmelt in shrub tundra is complicated by the emergence and spring-up of tall tundra shrubs which form a canopy over the snowpack changing the energy inputs to the underlying snow (Pomeroy et al., 2006; Bewley et al., 2010; Ménard et al., 2012). Meltwater infiltration into frozen soils can be restricted, limited, or unlimited depending on the degree of soil saturation by liquid water and ice and the over-winter development of ice layers on top of the soil (Janowicz, 2001; Janowicz et al., 2003). Frozen mineral soils usually have limited infiltration characteristics. The degree of saturation can be estimated from soil porosity, and the volumetric moisture content measured the preceding fall if overwinter soil moisture changes are minimal. Substantial mid-winter melts can create ice layers, which restrict infiltration such that most snowmelt forms overland flow (Gray et al., 2001). During summer, most rainfall infiltrates the porous soils and then is withdrawn by plant roots for evapotranspiration associated with the growth of mountain tundra and forests (Granger, 1999). Evapotranspiration occurs quickly from wet surfaces such as water bodies, wetted plant canopies, and wet soil surfaces and relatively slowly from unsaturated surfaces such as bare soils and plant stomata (Granger and Gray, 1989). Any hydrological model to be used for climate change assessment in this region must correctly address these hydrological processes.

To assess the impact of climate change and variability on the cold regions hydrological cycle, models require a full set of physically based representations of cold regions hydrological processes including direct and diffuse radiation to slopes, longwave radiation in complex terrain, intercepted snow, blowing snow, sub-canopy turbulent and radiative transfer, sublimation, energy balance snowmelt, infiltration to frozen and unfrozen soils, rainfall interception, evapotranspiration, sub-surface flow, depressional storage fill and spill, saturation excess overland flow and routing of surface, sub-surface, and streamflow. The Cold Regions Hydrological Modelling platform (CRHM) (Pomeroy et al., 2007) is a flexible model assembly system that is primarily physically based and represents all of the above-mentioned cold region hydrological processes. CRHM has been widely applied to studies in the continental climate of Western Canada in British Columbia, Yukon, the Canadian Rockies, the Canadian Prairies, Northwest Territories, and other cold regions in the Tibetan-Qinghai Plateau, Patagonia, the Pyrenees, and the Alps (Ellis and Pomeroy, 2007; Dornes et al., 2008; Ellis et al., 2010; López-Moreno et al., 2012; Fang et al., 2013; Zhou et al., 2014). CRHM was evaluated in the recent SnoMIP2 snow model intercomparison and performed relatively well in modelling forest snowmelt at sites in Switzerland, USA, Finland, and Japan (Rutter et al., 2009).

The difference between the spatial scales of climatic and hydrological models makes coupling the two sets of models for climate change analyses very challenging (Kite and Haberlandt, 1999). Hydrological modelling generally requires spatial resolutions of 0.1° latitude and longitude or smaller (Salathé, 2003). The resolution of a General Circulation Model (GCM), for instance the T63 version of the Environment Canada CGCM3.1 model, is 2.8°, which is much lower than needed for direct application in a semi-distributed hydrological model such as CRHM without downscaling. The GCM outputs also are not capable of capturing effects of the sub-grid features such as orography, cloud, convection, and land-cover. Bennett et al. (2012) assessed hydroclimatological modelling uncertainties in northwestern Canada and showed that uncertainties from GCMs, emissions scenarios and hydrological parameterizations were 84%, 58%, and 31%, respectively, for winter in the 2050s in headwaters of the Peace, Fraser and Campbell Rivers in northeastern British Columbia. Wilby and Harris (2006) found that simulations of low streamflows are very sensitive to uncertainties in GCMs. Wilby (2005) has recommended that sensitivity analyses can help in quantifying hydrological uncertainty from various sources in climate-change impact studies. One method for doing this is the delta method, which examines the change in a variable per degree Celsius of global warming (Fowler et al., 2007). While this method is very clear for hydrological response studies, the assumption of linear scaling of impact with temperature for non-linear variables such as precipitation and for extremes introduces uncertainties. An alternative method to delta method scaling in impact studies is the application of climate models projections to simulate future hydrology. This method can take advantage of a large number of simulated future climates calculated using various atmospheric and feedback assumptions (scenarios). However, projected changes in regional precipitation are very
uncertain in future climate scenarios due to coarse simulation of the synoptic dynamics that are responsible for most precipitation events (IPCC, 2014) and the problems of scale mismatch between hydrological and climate models as noted above.

In this study, an attempt is carried out to estimate components of the hydrological cycle using a physically based model created using CRHM that will in turn be able to assess the sensitivity of the hydrological response to climate change in a Yukon mountain basin. The model can project changes to snow dynamics, precipitation, and timing and magnitude of runoff. Given the uncertainties and to some extent disagreement between the climate model outputs for enhanced greenhouse gas scenarios over time, this paper examines the sensitivity of hydrological responses to changes in air temperature and precipitation by perturbing a recent measured time series of air temperature and precipitation up to the range of changes that are predicted by most climate models under late 21st century scenarios of global change. This linear sensitivity analysis is a step in exploring the possible alteration of the cold regions hydrological cycle in Yukon by climate change. The main purpose of this study is therefore to highlight the combination of changes in temperature and precipitation that are necessary to induce important changes in basin hydrology. Knowing what combination of warming and precipitation change can induce hydrological change in the North can help in assessing the threat to future water management from climate change.

METHODS

Study area and data sources

Wolf Creek Research Basin (WCRB) drains 179 km² of mountainous terrain in the headwaters of the Yukon River in Canada (Figure 1). The ecological diversity of the basin combined with the available long-term comprehensive hydrometeorological data (18 years) in each of three distinct ecosystems (boreal forest, shrub tundra, and alpine tundra) makes WCRB a suitable subject for this type of climate change study. The elevation of the basin ranges from 660 to 2080 m with a generally north-easterly aspect. Studies have shown slope and aspect to be important controls on runoff processes (Woo and Carey, 1999; Pomeroy et al., 2003). The hydrology of WCRB is strongly influenced by long winters (usually eight months) with small groundwater contributions to streamflow, spring melt of the seasonal snowpack causing high streamflow, and low streamflow in the relatively short, warm, and dry summer (McCartney et al., 2006). Streamflow at the outlet of Wolf Creek is typical of a mountainous subarctic regime with snowmelt-dominated peak flows in late May or early June and low flows in March (Janowicz, 1999). Along the main branch of Wolf Creek is Coal Lake (area 1 km²), which controls winter baseflows. At low elevations, jack pine, white and black spruce, and trembling aspen forest stands predominate (Francis et al., 1999). Above the treeline, shrub tundra with birch and willow shrub heights from 30 cm to 2 m occupies the majority of the basin (58%). At the highest elevations is an alpine tundra ecozone of bare rock and short tundra moss and grass vegetation.

The near-surface geology of WCRB consists primarily of sedimentary rocks such as limestone, sandstone, siltstone, and conglomerate capped by a mantle of till at least 2 m thick, as well as glaciofluvial and glaciolacustrine deposits (Seguin et al., 1998). Soils are generally well drained due to the coarse parent materials. Lower elevation forest soils are generally of clay to gravel texture, while upland soils are primarily of sandy loam to gravely sandy loam texture. Surface organics commonly extend to depths of 0.1 m and are deepest in riparian areas and north-facing slopes (Zhang et al., 2010). Estimations by Lewkowicz and Ednie (2004) suggest that 43% of the basin contains permafrost which restricts the movement of water beneath the surface, particularly in the case of saturated frozen soils (Carey and Woo, 2001; Quinton et al., 2009).

Hydrometeorological data have been collected from the water year of 1993–1994 to 2010–2011 from three main meteorological stations in WCRB, one in each primary ecozone, and from four streamflow gauges (Figure 1, Table I). Hourly air temperature, relative humidity, wind speed, incoming shortwave radiation, and daily precipitation observations from above-canopy meteorological stations and streamflow data from hydrometric stations are used in this study. Precipitation was measured by tipping bucket rain gauges, unshielded ‘BC style standpipe’ precipitation gauges, and Nipher-shielded Meteorological Service of Canada (MSC) snowfall gauges. Nipher gauge solid precipitation measurements were corrected using a wind undercatch correction equation (Goodison et al., 1998) with wind speeds measured from nearby gauge-height anemometers. Gaps in data were infilled by establishing regression equations for meteorological variables between each of the three meteorological stations and the Whitehorse Airport MSC station located 15 km from WCRB. Because the basin is relatively small (∼179 km²) and stations are close to each other, meteorological variation is largely controlled by orography and hence changes in elevation rather than distance between stations. Measurements of snow depth and density along snow survey transects were collected at least monthly by Yukon Environment and university researchers at each of the three meteorological stations. These measurements and snow water equivalent (SWE) measured at a snow pillow located at the subalpine site
provide model diagnostic information for each ecozone (Pomeroy and Gray, 1995; Pomeroy et al., 1999).

Cold regions hydrological model for WCRB

Models created using CRHM can have a wide variety of structures with differing levels of process detail and representation including structures that are suitable for a northern basin such as WCRB (Pomeroy et al., 2007). The CRHM system is very flexible and creates ‘objected-oriented’ models for particular basins, environments, and predictive needs. Recent developments include options for redistribution of alpine blowing snow (MacDonald, 2010), an improved physical basis to soil moisture accounting, fill-and-spill depressional storage (Fang et al., 2010), and enhanced forest canopy interception and radiation modules (Ellis et al., 2010). For this study, a CRHM model that accurately characterized the surface and near-surface cold regions hydrological processes of WCRB was needed.

A set of physically based process modules describing the major processes can be combined into a model informed by results from previous modelling experiments in the research basins such as WCRB (Pomeroy and Granger, 1999; Pomeroy et al., 2003, 2006; McCartney et al., 2006; Carey et al., 2007, Dornes et al., 2008; Quinton and Carey, 2008, MacDonald et al., 2009).
Modules were selected that could be run to robustly simulate the hydrological cycle of the region in a primarily physically based manner. Figure 2 shows the schematic setup of these modules, which include:

1) Observation module: reads and adjusts the meteorological data with hydrological lapse rate, elevation, and wind-induced undercatch.
2) Radiation module: computes the theoretical global radiation, direct and diffuse solar radiation, and maximum sunshine hours based on latitude, elevation, ground slope, and azimuth (Garnier and Ohmura, 1970).
3) Sunshine hour module: estimates sunshine hours from incoming short-wave radiation and maximum sunshine hours.
4) Slope radiation module: adjusts the short-wave radiation on the slope with incoming short-wave radiation on a level surface.
5) Long-wave radiation module: calculates incoming long-wave radiation using short-wave radiation (Sicart et al., 2006).
6) Albedo module: estimates snow albedo throughout the winter and into the melt period and also indicates the beginning of melt (Verseghy, 1991).
7) Canopy module: estimates the snowfall and rainfall intercepted by and sublimated or evaporated from the forest canopy and unloaded or dripped from the canopy, updates sub-canopy snowfall and rainfall, and calculates short-wave and long-wave sub-canopy radiation and turbulent transfer to snow. This module has options for open, small forest clearings, and forested environment (Ellis et al., 2010; 2013).
8) Blowing snow module: simulates the inter-hydrological response unit (HRU) wind redistribution of snow transport and blowing snow sublimation losses throughout the winter period (Pomeroy and Li, 2000).
9) Energy balance snowmelt module: estimates snowmelt by calculating the energy balance of radiation, sensible heat, latent heat, ground heat, advection from rainfall, and change in internal energy (Marks et al., 1998).
10) All-wave radiation module: calculates the net all-wave radiation from short-wave radiation for snow-free conditions (Granger and Gray, 1990).
11) Snowmelt infiltration module: This module has two algorithms for infiltration into frozen and unfrozen soils. One estimates snowmelt infiltration into frozen soils using a parametric equation describing the results of a finite difference heat and mass transfer model (Gray et al., 2001) and the other one estimates rainfall infiltration into unfrozen soils with macropores using a soil classification (Ayers, 1959). Surface runoff forms when snowmelt or rainfall exceeds the infiltration rate.
12) Evaporation module: Granger’s evapotranspiration expression (Granger and Gray, 1989; Granger and Pomeroy, 1997) estimates actual evapotranspiration from unsaturated surfaces using an energy balance and extension of Penman’s equation to unsaturated conditions; the Priestley and Taylor evaporation expression (Priestley and Taylor, 1972) estimates evaporation from saturated surfaces such as stream channels and lakes. Both evaporation algorithms modify moisture content in the canopy interception store, ponded surface water store, and soil column and by linking to the soil module are restricted by a soil water withdrawal relationship to ensure continuity of mass. The Priestley and Taylor evaporation also updates moisture content for saturated surfaces.

Table I. Description of the main meteorological and hydrometric stations within Wolf Creek

<table>
<thead>
<tr>
<th>Station ID</th>
<th>Station ID</th>
<th>Station ID</th>
<th>Station ID</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Wolf Creek</td>
<td>Coal Lake Outlet</td>
<td>Granger Creek</td>
<td>Alaska Hwy</td>
</tr>
<tr>
<td>29AB006</td>
<td>29AB005</td>
<td>29AB007</td>
<td>29AB002</td>
</tr>
<tr>
<td>Location</td>
<td>Location</td>
<td>Location</td>
<td>Location</td>
</tr>
<tr>
<td>60° 29.45′ N, 135° 17.50′W</td>
<td>60° 30.61′ N, 135° 9.74′W</td>
<td>60° 32.79′ N, 135° 11.08′W</td>
<td>60° 36′ N, 134° 57′W</td>
</tr>
<tr>
<td>Drainage area (km²); record period</td>
<td>Drainage area (km²); record period</td>
<td>Drainage area (km²); record period</td>
<td>Drainage area (km²); record period</td>
</tr>
</tbody>
</table>
13) Soil and hillslope module: estimates soil moisture balance, depressional storage, surface/sub-surface flows in two soil layers and groundwater discharge in a groundwater layer, and interactions between surface flow and groundwater (Leavesley et al., 1983; Pomeroy et al., 2007; Dornes et al., 2008; Fang et al., 2010, 2013). The top layer is recharge layer, which receives infiltration from depressional storage, snowmelt, and rainfall. Evaporation takes water first from canopy interception and depressional storage and then from the recharge layer or from both soil column layers via transpiration based on the plant available soil moisture and the rooting depth (Armstrong et al., 2010). Aquifer recharge occurs via percolation from the lower soil layer or directly from the surface through macropores. Horizontal and vertical drainage from the soil and groundwater layers is regulated by Darcy’s flux, parameterized using Brooks and Corey’s (1964) relationship to estimate the actual hydraulic conductivity in unsaturated zone.

14) Routing module: Clark’s lag and route timing estimation method is used to route runoff (Clark, 1945).

The proposed physically based model operates on the spatially distributed control volumes of the HRU which have been found useful for modelling in basins where there is a good understanding of hydrological behaviour, but incomplete detailed sub-surface information to permit a fully distributed fine scale modelling approach (Dornes et al., 2008). HRUs in a given basin are segregated based on the land-cover, slope, aspect, and elevation utilizing a digital elevation model (DEM). The temporal resolution of this CRHM model is hourly, while the spatial resolution is that of the HRU which vary from 0.92 km² to 25.4 km² in WCRB.

Model parameter estimation

Parameters were estimated from research basin results and those from other mountain modelling studies (Smoky River, Marmot Creek in Canada; Pomeroy et al., 2011; 2013) with limited calibration for timing and routing of the streamflow. Blowing snow fetch distances of 500 m were found from the study of MacDonald et al. (2009) and used for the alpine and shrub tundra HRUs. Vegetation heights determined by field surveys were used for the model simulations. Values of the blowing snow redistribution factor were chosen as 2 and 5 for the alpine and shrub tundra HRUs, respectively (MacDonald et al., 2009). Blowing snow was inhibited for the forest HRU. Initial soil saturation prior to the snowmelt infiltration was estimated from both pre-melt volumetric soil moisture content observations and soil porosity measurements. The yearly antecedent soil moisture was measured by water content reflectometer measurements at each station. Soil types (organic and mineral soils) and their porosity values were measured by Carey and Woo (2005), Quinton et al. (2005) and Zhang et al. (2010). Initial soil temperature was measured by soil thermocouples prior to the major snowmelt. The environment coefficient and surface saturation for frozen soil
infiltration were set up based on the recommended values from Gray et al. (2001). Infiltration opportunity time was calculated by model simulation of snowmelt duration. Previous studies on soil properties in the Wolf Creek (Carey and Woo, 2001; Carey and Quinton, 2004; Carey and Woo, 2005; Quinton et al., 2005) found that the active soil layer was composed of an upper organic soil layer over a mineral soil layer. Measured depth and porosity values for both organic and mineral soil layers were used to approximate the capacity of the soil column layer. The soil recharge zone, a top layer of the soil column, was assumed to be shallow and initially saturated.

Modelling the complex processes controlling streamflow generation in the model was relatively difficult due to the varying magnitude and timing of the hydrological processes over short distances, aufeis formation, river ice formation and breakup, snow dammed channels, and the presence of discontinuous permafrost slopes (Carey and Woo, 2001; McCartney et al., 2006; Boucher and Carey, 2010). Lag and route times in channels between sub-basins were not measured by any study in Wolf Creek. The capacity of surface depressional storage was also unknown and so both were calibrated. Subsurface drainage factors were thought to be affected by subsurface flow perched over thawing frozen soil which is not yet fully parameterized in CRHM and so were calibrated from streamflow. Table II shows the routing and drainage parameters calibrated from measured Wolf Creek streamflow at the Alaska Highway using the dynamically dimensioned search (DDS) approach (Tolson and Shoemaker, 2007) and a calibration based upon 1000 model runs over 3 years (1998–2001) for which measurement confidence is relatively high. The DDS approach is a heuristic search method developed to obtain an acceptable optimal solution within a given number of model evaluation runs. First, the DDS algorithm searches the parameter set globally, and then it looks for local solutions as the number of model evaluations approaches the specified try number. The transition from global to local search is carried out dynamically by reducing the size of the neighbourhood or dimensions of the model parameters tuned. The calibrated parameters are the lag and route timing and storage constant, sub-surface drainage rates and depressional storage of the HRUs. The range of the parameters was selected so that they were reasonable for the study area based on the experience of the authors; the DDS search algorithm then optimized the parameters within this range. It is intended that more physically accurate parameterisations of subsurface drainage will become available, based on the studies by Quinton et al. (2005), but until that time, calibration of the subsurface drainage is needed as per the approach of Dornes et al. (2008).

Model spatial configuration

A spatially distributed modeling structure (Figure 3) was developed with five sub-basins and 29 HRUs in Wolf Creek basin, based on three elevation bands, four groupings of slope and aspect and three land cover ecozones (Figure 1). A DEM with 30 m cell resolution was prepared by Dr. Lawrence Martz (Dept of Geography, University of Saskatchewan). The extracted elevation, aspect, and slope were then intersected with the basin land cover feature in ArcGIS, which generates the HRUs based on elevation, aspect, slope, and land cover (Figure 1). All the HRUs were categorized into one of three ecozone groups (Janowicz, 1999); alpine tundra, shrub tundra, or forest for parsimonious parameterization. Four streamflow gauges along Wolf Creek serve as the outlet of sub-basins or the whole basin. An additional sub-basin is Mid-Wolf Creek between Coal Lake and the forested ecozone in the lower basin. CRHM modelling structures were replicated by ecozone to simulate the hydrological processes in the five sub-basins. A set of physically based modules were assembled for a number of HRUs and the structure was replicated for each sub-basin. Each sub-basin then possessed the same module configuration but was permitted varying parameter sets and varying numbers of HRUs. Streamflow from each sub-basin was routed along the main channel of Wolf Creek.

Sensitivity analysis

After defining the model structure, parameters, and the control volumes of computation on the CRHM platform, a detailed sensitivity assessment of hydrological response to climate warming at the given basin could be carried out.

Table II. List and range of the lag and route timing and drainage parameters calibrated using the DDS algorithm in channels and HRUs

<table>
<thead>
<tr>
<th>Parameter name</th>
<th>Unit</th>
<th>Lower bound</th>
<th>Upper bound</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basin-scale groundwater lag time</td>
<td>h</td>
<td>0</td>
<td>100</td>
</tr>
<tr>
<td>Basin-scale runoff lag time</td>
<td>h</td>
<td>0</td>
<td>100</td>
</tr>
<tr>
<td>Groundwater storage constant</td>
<td>day</td>
<td>0</td>
<td>50</td>
</tr>
<tr>
<td>Groundwater lag time</td>
<td>h</td>
<td>0</td>
<td>100</td>
</tr>
<tr>
<td>Drainage rate from groundwater</td>
<td>mm/day</td>
<td>0</td>
<td>0.5</td>
</tr>
<tr>
<td>Aggregated storage constant</td>
<td>day</td>
<td>0</td>
<td>50</td>
</tr>
<tr>
<td>Aggregated lag time</td>
<td>h</td>
<td>0</td>
<td>100</td>
</tr>
<tr>
<td>Surface runoff storage constant</td>
<td>day</td>
<td>0</td>
<td>50</td>
</tr>
<tr>
<td>Surface runoff lag time</td>
<td>h</td>
<td>0</td>
<td>100</td>
</tr>
<tr>
<td>Drainage rate from depressional</td>
<td>mm/day</td>
<td>0</td>
<td>0.5</td>
</tr>
<tr>
<td>storage to groundwater</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
with climate change impact model experiments assessing: the effect of temperature and/or precipitation forcing change on hydrological response assuming that the basin vegetation and permafrost do not change. The purpose of this methodology is to look at first-order impacts of climate change on hydrological processes including snow regime and runoff magnitude and timing. This methodology is a first step examination of response to changing climate before comprehensive transient impacts due to concurrently changing vegetation and permafrost can be studied. For instance, Ménard et al. (2014) showed for snowmelt in a small portion of the catchment that mountain topography can reduce the impact of changing shrubs on snow redistribution, spring energy balance, and meltwater generation. In future work, second-order impacts of changing climate such as transient changes in vegetation and permafrost on the hydrology of WCRB will be investigated.

The range of the changes in precipitation and warming considered for this study is partly based on the changes estimated by Representative Concentration Pathways (RCPs) which are an alternative to the conventional climate change scenarios (Moss et al., 2010) and were used in the recent Fifth Assessment Report (AR5) of the Inter-governmental Panel on Climate Change (IPCC, 2014). Four new atmospheric composition pathways corresponding to specific radiative forcing values of 2.6 to 8.5 W/m² were used as a basis for long-term and near-term modeling experiments in climate change studies. Based on RCP2.6 and RCP8.5 pathways, respectively, a warming of up to 2 °C with an increase in annual precipitation of less than 10% and a warming of up to 5 °C with a 20% increase in annual precipitation are expected for the southern Yukon region around Wolf Creek. Most modelled scenarios project the future climate to be wetter, but there are some scenarios used in AR5 that show a regional decrease in precipitation up to 15% for the 2080s. As hydrological impacts in a northern basin are largely driven by changes to winter and spring climate conditions and modelled scenarios differ in their assessment of seasonality of climate change, seasonal variations in temperature and precipitation were considered unnecessary for an initial climate sensitivity study. Based on this consideration and the range of climate scenarios reviewed, a temperature increase of up to 5 °C and a range of −20% to +20% change in precipitation are analysed in this paper.

RESULTS

Snow regime simulation

As snowmelt is the major runoff-producing event of the year and streamflow forecasts are derived from seasonal maximum snow accumulation estimates, simulating snow dynamics correctly is critical for hydrological impact assessments in Yukon. The CRHM model of WCRB was evaluated against snow surveys at the three meteorological stations (corresponding to the three ecozones) and against basin streamflow at the outlet. The purpose of this
evaluation was to ascertain whether the model had sufficient physical realism in its predictions to be used for climate change assessment, not to assess it as a streamflow prediction tool. Modelled snow accumulation for each HRU was aggregated by the ecozone-weighted average (alpine, shrub tundra, and forest) and then compared to the snow survey measurements of SWE (Figure 4) made by Yukon Environment—Water Resources and its federal predecessors. The modelled snow accumulation is shown as SWE and is a result of snowfall, rainfall, blowing snow transport, sublimation, snow interception, canopy unloading, evaporation, and snowmelt calculations and so assesses a major part of the model and over half of the hydrological year.

Assessment of model performance for snow accumulation used the root mean square error (RMSE) and the mean absolute error (MAE)—both in mm SWE. RMSE ranged from 35 mm in the alpine to 29 mm in the forest and MAE ranged from 28 mm in the alpine to 23 mm in the forest. These statistics indicate model errors of less than 25% in estimating snow dynamics. As shown in Figure 4, the model often overestimated snow accumulation at all elevations compared to snow surveys. The representativeness of the snow surveys and their accuracy must be considered in assessing this. Errors in weighted snow tube measurements of SWE of 10% are common (Pomeroy and Gray, 1995) when compared to snow pit gravimetric measurements of density and depth. As well, the alpine and shrub tundra snow surveys did not include a drift accumulation zone. Drift accumulation zones are very important to the basin-scale snow mass balance of Wolf Creek (Pomeroy et al., 1999; 2003). Hence, the alpine and shrub tundra modelled snow accumulation will likely exceed the measured when the measurements do not include drift areas. Overestimation of snow accumulation in the forest may be contributed to by differences between the aggregated forest HRU discontinuous canopy structure and that of a single snow survey transect in a continuous white spruce forest near the forest meteorological station. The snow survey line in the forest may represent a more dense forest canopy than is typical of the aggregated forest HRUs. Overall, the model seemed capable of simulating the SWE regime and the timing of peak snow accumulation and initiation of snowmelt well.

Streamflow simulation

The model was calibrated against the gauged flows of Wolf Creek at the Alaska Highway for the period of 1998–2001 (Figure 5). To assess the multiple year performance of the model, flows were simulated from 1993 to 2011 and compared to streamflow observations (Figure 5). The RMSE, MAE, and Nash–Sutcliff statistic (NS) for the calibration period are 0.72 m$^3$/s, 0.44 m$^3$/s, and 0.49 and for the full 18-year hourly simulation are 0.88 m$^3$/s, 0.49 m$^3$/s, and 0.15, respectively. The error range and bias are similar between calibration and simulation periods, but the NS statistic degrades, showing weaker hydrograph simulation over the simulation period. The acceptable performance of the model for both ecozone-scale SWE and basin-scale streamflow simulations given the minimal calibration of model parameters, along with the strong physical basis of its structure, suggests that this model can be used for climate change sensitivity simulations of snowpack and streamflow.

![Figure 4. Comparisons of simulated and observed snow water equivalent in the three ecozones of WCRB: alpine, shrub tundra, and forest](image-url)
Impact of changing temperature and precipitation on snow regime

The sensitivity analysis involved raising air temperatures by one-degree increments for each hourly interval up to 5 °C and then altering daily precipitation either upwards or downwards by a multiplier for ±10% and ±20%. The impact of raising hourly air temperatures by 1 °C increments up to 5 °C on the seasonal snow regime is shown in Figure 6 for the alpine, shrub tundra, and forest ecozones from October 1993 to September 2011. Climate warming of up to 5 °C decreased the simulated annual peak snowpacks from 30% to 45%. There are substantial declines in snow accumulation and duration of snow season with increasing temperature over most winters, with some seasons (e.g. 2002–2003) showing large sensitivities and some (e.g. 1997–1998) almost no sensitivity to temperature increases. Warming affects the phase of precipitation, causing a shift from snowfall to rainfall in the spring and fall transition seasons, and so causes the snow season to start later and end earlier. It also can accelerate the initiation of snowmelt, which in some years slowed the melt rate as the melt period was shifted forward into a lower solar irradiance period. Despite this effect, in most years snowmelt ended earlier as temperatures increased.

The peak SWE was calculated as the maximum SWE over the year; this generally occurs in April or May of each calendar year. The peak SWE and how it is affected by climate warming is shown in Figure 7 for the shrub tundra ecozone. In some years, there are very strong decreases in peak SWE with increasing temperature (e.g. 2000, 2005, 2006), whereas in other years there is very little decrease (1998, 2002, 2007). The most sensitive years are the wetter years as these often include late spring snowfall that occurs under relatively warm conditions and so is most vulnerable to phase change with warming.

A sensitivity analysis with warming air temperatures (up to 5 °C) and both increasing (+10%, +20%) and decreasing (−10%, −20%) precipitation was performed to assess whether changes in precipitation would dampen

Figure 5. Observed (points) and simulated (line) streamflow discharge of Wolf Creek at the Alaska Highway, Oct 1993 to Sep 2011. The period of 1998–2001 was used for calibration of parameters listed in Table II

Figure 6. Snow accumulation regimes for alpine, shrub tundra, and forest ecozones, under current climate and incremental warming of hourly air temperatures by up to 5 °C
or magnify the effect of warming climate on WCRB. The overall response of peak SWE magnitude and timing in each ecozone to changing precipitation with warming is shown in Figure 8. Peak SWE declined markedly with increasing temperature and decreasing precipitation in all ecozones, with a slightly stronger response to precipitation in the alpine, a slightly stronger response to temperature in the shrub tundra and a fairly equal response to both temperature and precipitation in the forest ecozone. The response of peak SWE to changing precipitation with moderate warming (2 °C) was non-linear suggesting sensitivity to the combination of warming and drying and relative insensitivity to warming and wetting. For instance, with 2 °C warming, a 20% decrease in precipitation caused a 35% reduction in peak SWE, and a 10% decrease in precipitation caused a 26% reduction, but a 20% increase in precipitation caused only a 9% increase in peak SWE. The timing of peak SWE advanced dramatically with increases in temperature and was only impacted by increases or decreases in precipitation in the forest and shrub tundra ecozones. Besides a general advance in the date of peak SWE with combined warming and drying, there was no clear pattern in the alpine ecozone due to the complex interaction of warming and drying.
blowing snow transport from the alpine snowpack with air temperature and precipitation, suggesting that the timing of peak SWE in the alpine is insensitive to the most likely scenarios of warming and wetting climate.

For increased precipitation without warming, the peak SWE responded linearly in each ecozone with a 22% increase for a 20% precipitation increase. The shrub tundra snow regime showed the greatest sensitivity to the simultaneous drying and warming; a 20% decrease in precipitation along with 5 °C warming resulted in a 52% decline in SWE (from 162 mm to 84 mm) and a 25-day advance in timing. The high sensitivity of the shrub tundra snow regime to warming temperatures is contributed to by the diminished redistribution of blowing snow from alpine to shrub tundra as the temperature warms (Figure 9). The relatively low sensitivity of the alpine peak SWE date to temperature is due to the colder climate at high elevations. Warming temperatures in the relatively cold alpine zone cause only a small reduction in snowfall but can cause a substantial reduction in blowing snow transport from the alpine zone. Reduced snowfall and reduced snow erosion compensate for each other to some degree, but compensation will vary with the meteorological characteristics of each snowstorm and snowfall event. The relatively lower sensitivity of the forest peak snow accumulation to warming is due to the temperature-sensitive intercepted snow unloading processes in forests counteracting reduced snowfall at higher temperatures; warmer temperatures lead to greater unloading of intercepted snow and lesser sublimation loss (Gelfan et al., 2004; Figure 9). A fundamental question regarding the impact of climate change on snowpack is whether the maximum likely increase in precipitation of 20% can compensate for the reduced peak SWE caused by the maximum likely warming. This is explored in Figure 10 where the cumulative distribution of peak SWE in each ecozone shows that the impact of a 20% increase in precipitation can compensate for that of 5 °C of warming on peak SWE only in the highest 1–2% of snowfall years—for lower snowfall years it is less likely that increases in precipitation can compensate for the impacts of increased warming on peak SWE.

Table III summarizes the basin scale magnitude and timing of the peak SWE, snow season initiation, end and length, peak flow, and total annual runoff for current climatic conditions and for scenarios of warming and changes in precipitation. The average peak SWE in the basin declined (55%) from 136 mm to 61 mm with 5 °C warming and a 20% decline in precipitation and increased (24%) to 169 mm without warming and with 20% increase in precipitation. The most likely scenario of 5 °C warming and 20% increased precipitation resulted in 18% decrease in annual peak SWE. With 5 °C warming and no changes in precipitation winters in WCRB are predicted to begin 17 days later and end 37 days earlier than current winters; a concomitant 20% increase in precipitation would lengthen the winter season by only one week. With warming of only 2 °C, increased precipitation of as little as 10% was able to compensate for the effect of warming and further increases in precipitation resulted in increased peak SWE. With 5 °C warming, peak snowpack decreased in all scenarios, even with increased precipitation of 20%, which is consistent with the frequency distribution results from Figure 10.

Impact of changing temperature and precipitation on streamflow

Streamflow simulations are sensitive to changes in the whole range of hydrological processes and water balance components as integrated by the basin. The daily runoff frequency curves for low and high runoffs of Wolf Creek (at the Alaska Highway) are shown in Figure 10. Streamflow exhibits a very different response to the changes in precipitation under warming from that of the historical snowmelt driven regime. Under all temperature increases concomitant with precipitation increases of 20%, medium and low flows in Wolf Creek are greater than those the current regime. This persists for high flows as well, except with 5 °C of warming, for which low flows increased but high flows decreased compared to runoff under the current climate; this dampening of hydrological variability under a very likely climate scenario is of great importance to water resources management. The overall change in annual runoff over the 18 years of simulation is shown in Figure 11. Change in runoff contrasts with the change in peak SWE (Figure 8) in that runoff is much more sensitive to precipitation than to temperature change. A 1 °C increase in temperature resulted in a 4%
decrease in annual runoff and grew to a 14% decrease for a 5 °C temperature increase. The combination of 5 °C warming and 20% decreased precipitation reduced annual runoff to 96 mm/year, 44% less than the current value (171 mm/year, Figure 11); however, 5 °C with 20% increased precipitation increased runoff by 20% to 205 mm/year. The sensitivity of runoff to temperature is because of the longer snow-free season and increased opportunity and energy for evapotranspiration with increasing temperature. This is illustrated by the 45-mm increase in annual actual evapotranspiration with 5 °C of temperature rise (Figure 9). It is interesting to note that an 8% increase in precipitation is necessary to compensate for the impact of 5 °C warming on runoff from Wolf Creek.

Figure 12 shows that peak runoff rates increase proportionately greater than do changes in precipitation, demonstrating a high sensitivity of runoff to changes in precipitation. Peak runoff timing advanced under most combinations of changed precipitation and warming (Figure 12). Peak runoff was more sensitive to temperature changes than was annual runoff. For the combination of 2 °C warming and 20% increased precipitation, runoff timing advanced 4 days and peak rate increased by 11%; however, for 5 °C warming and 20% increased

Table III. The area weighted average of SWE and discharge at the Wolf Creek outlet and their different timing in the hydrological year starting 1 October and ending 30 September for different scenarios of climate warming and changes in precipitation

<table>
<thead>
<tr>
<th>Variable</th>
<th>Index</th>
<th>Unit</th>
<th>T:0 °C</th>
<th>T:2 °C</th>
<th>T:5 °C</th>
<th>T:0 °C</th>
<th>T:120%</th>
<th>T:0 °C</th>
<th>T:80%</th>
<th>T:120%</th>
</tr>
</thead>
<tbody>
<tr>
<td>SWE</td>
<td>Peak</td>
<td>mm</td>
<td>135.9</td>
<td>117.3</td>
<td>84.8</td>
<td>169.4</td>
<td>61.0</td>
<td>107.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Accumulation start</td>
<td>Day of year</td>
<td>5</td>
<td>7</td>
<td>22</td>
<td>4</td>
<td>27</td>
<td>18</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Timing of peak</td>
<td>Day of year</td>
<td>167</td>
<td>163</td>
<td>154</td>
<td>172</td>
<td>148</td>
<td>155</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Timing of snow free</td>
<td>Day of year</td>
<td>271</td>
<td>254</td>
<td>234</td>
<td>273</td>
<td>230</td>
<td>237</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Length of snow season</td>
<td>Day</td>
<td>265</td>
<td>248</td>
<td>212</td>
<td>269</td>
<td>202</td>
<td>219</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Discharge</td>
<td>Annual runoff</td>
<td>mm</td>
<td>171</td>
<td>160</td>
<td>147</td>
<td>235</td>
<td>95</td>
<td>205</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Annual peak rate</td>
<td>m^3/s</td>
<td>4.1</td>
<td>3.6</td>
<td>2.9</td>
<td>5.3</td>
<td>2.1</td>
<td>3.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Timing of rising flow</td>
<td>Day of year</td>
<td>160</td>
<td>139</td>
<td>119</td>
<td>161</td>
<td>94</td>
<td>119</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Timing of peak flow</td>
<td>Day of year</td>
<td>263</td>
<td>244</td>
<td>236</td>
<td>265</td>
<td>232</td>
<td>235</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Return to baseflow</td>
<td>Day of year</td>
<td>353</td>
<td>337</td>
<td>343</td>
<td>357</td>
<td>340</td>
<td>336</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Summer flow length</td>
<td>Day</td>
<td>193</td>
<td>197</td>
<td>225</td>
<td>196</td>
<td>246</td>
<td>217</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
precipitation timing advanced 27 days and peak rate declined by 10%. This suggests that increased and earlier high flow potential is possible with moderate warming and maximum wetting but not with maximum warming which produces much earlier peak flow timing and reduced peak flows. Drying under 2°C warming caused a substantial reduction in peak flow; precipitation decreases (10–20%) caused decreases in peak runoff rates (25–35%). Overall, increases in temperature tended to reduce peak flows by desynchronizing melt through accelerating the timing of spring flows (Table III) and reducing summer and fall flows as actual evapotranspiration increases in the longer snow-free period. The impact on peak flow rates of warming by 2°C and 4°C can be compensated by increases in precipitation of 10% and 20%, respectively; however, even with a 20% increase in precipitation, peak flow rates decreased as warming increased from 4°C to 5°C.

**Synthesis**

The average annual elevation-corrected precipitation of WCRB is 334 mm over 1993–2011, suggesting that runoff efficiency would only decline from 0.51 to 0.45 with 5°C of warming. Conversely to this modest temperature sensitivity,
runoff was more sensitive than the snow regime to precipitation change and the impact of up to 5 °C of warming on annual runoff could be compensated for by an increase in precipitation of less than 8%. The ambiguity in the direction of change in precipitation causes great uncertainty in the future hydrology of WCRB. If a 5 °C temperature increase is assumed, then this could cause a 44% decrease in runoff if precipitation declines by 20%, a 14% decrease if precipitation does not change, and a 20% increase in runoff if precipitation increases by 20%. The combination of climate warming and decreased precipitation could cause dramatic declines in runoff, but if precipitation increased as many scenarios suggest, then there could be some compensation. For instance, for the not-unlikely scenario of precipitation increasing by 20% with 4 °C of warming, then runoff increases 23% and peak runoff rates are unchanged, although the timing of peak runoff is advanced by 20 days.

The changes in Wolf Creek streamflow are largely caused by the consequences of the decrease in depth and advance in timing of peak snowpack with climate change. The combination of 5 °C warming and 20% decreased precipitation cause modelled declines in peak SWE of 55% and the snow-covered period of 24%, resulting in increased evapotranspiration of 30% and decreased in annual runoff and peak runoff rates of 42% and 44%, respectively. This suggests that WCRB hydrology has a strong sensitivity to a ‘loss of cold’ that is connected to large decreases in snowpack with warming temperatures. The greatest hydrological sensitivity and potential for decline with increasing temperature is for peak snow accumulation, peak runoff, and their timing. The proportional drop in peak snow accumulation was closely reflected in the proportional drop in annual runoff, highlighting the importance of correctly estimating changes in snowpack when assessing changing hydrology in the region.

CONCLUSIONS

WCRB is Yukon’s most intensively studied headwater basin and has an 18-year archive of high altitude weather, snowpack, soils, and streamflow data that has been collated and corrected. A comprehensive cold regions hydrological process model developed using the CRHM Platform provides the basis for water balance estimates and climate sensitivity studies in a characteristic Yukon headwaters basin. Using parameters derived from the results of local scientific research, the model was able to adequately predict snow accumulation and melt in alpine, shrub tundra and forest zones, and basin streamflow over the simulation period. Perturbing the model with increases in hourly air temperatures from 1 to 5 °C, and both increases (to +20%) and decreases (to –20%) in daily precipitation provided the basis for a hydrological sensitivity study to possible ranges of climate change. The simulated changes show that snowpacks are very sensitive to warming in wet years and less sensitive in drier years; this is mostly due to spring contributions to wet year precipitation and the phase sensitivity of relatively warm spring snowfall (Harder and Pomeroy, 2014). On average, a 5 °C warming with no change in precipitation caused a 38% decline in maximum snow accumulation. The impact of 2 °C warming on snow could be fully compensated for by precipitation increasing by 20%, but greater warming (>3 °C) cannot be compensated with precipitation increases of this magnitude. If precipitation decreases with warming, the impacts on snowpack multiply and a decline of 55% of peak SWE could occur with both 5 °C warming and 20% decline in precipitation; though this is not a likely climate scenario, it would have catastrophic implications for winter ecology and snow-based transportation. The smaller snowpacks and warmer weather would cause an increase in the snow-free period by 52 days. These changes would also expand the evapotranspiration season, increasing annual evapotranspiration loss by 30% and the importance of rainfall–runoff mechanisms.

The impacts from 5 °C warming with a 20% decrease in precipitation on the hydrology of Wolf Creek would be very severe and result in snow accumulation, streamflow volume and peak runoff dropping approximately in half, peak flow timing advancing by one month and the snow-free period lengthening by 2 months. This contrasts strongly with the attenuated impact from the same warming and an increase in precipitation of 20% in which snow accumulation decreases by 21%, streamflow volume increases by 20% and peak runoff decreases slightly whilst the snow-free period lengthens by 46 days. These results show that while the impacts of warming on cold regions hydrological processes are unequivocal—reduced snow contribution to streamflow, shorter snow-covered period, and greater evapotranspiration; the magnitude and direction of the impact of warming on streamflow hydrology will depend on changes to precipitation. Therefore, both warming and changes to precipitation must be considered to evaluate future hydrology in Yukon. The model developed here can be applied to other Northern regions to assess the sensitivity to warming in environments with different climate, and vegetation conditions. It is considered most reliable for its treatment of climate change impact on snow hydrology. However, improvements to the treatment of frozen ground and inclusion of the transient response of permafrost and vegetation to changing climate would add great insight to the impact of local feedbacks to the climate system on hydrology. The extension of the
sensitivity analysis to seasonal temperature changes and changes in the intensity, duration, and frequency of precipitation as well as to its seasonality would reduce the uncertainty due to precipitation in the results. In the meantime, the results of this study may inform the early development of adaptation strategies and options which may include improving flood forecasting and warning systems, infrastructure design modification, land zonning changes, and changes to policy, regulation, and legislation.

ACKNOWLEDGEMENTS

The authors wish to acknowledge two decades of Wolf Creek Research Basin operation with substantial contributions to installation, maintenance and operation of the core snow surveys, meteorological stations and hydrometric stations under inclement conditions from Glenn Ford, Glen Carpenter, Kerry Paslowski, Martin Jasek, Dell Bayne, Newell Hedstrom, Raoul Granger, and Richard Essery amongst many others. Modelling support from Tom Brown and Xing Fang are greatly appreciated. Funding for the basin has come from Yukon Environment, Environment Canada, DIAND, NSERC, CFCAS, NERC, NOAA, and other sources. Funding for this study came from AAND Canada, Yukon Environment, and NSERC.

REFERENCES


HYDROLOGICAL SENSITIVITY OF A NORTHERN MOUNTAIN BASIN TO CLIMATE CHANGE


Tohola BA, Shoemaker CA. 2007. The dynamically dimensioned search (DDS) algorithm for computationally efficient watershed model


