

**DEVELOPMENT, EVALUATION, AND APPLICATION OF DOMINANT RUNOFF
GENERATION PROCESSES IN HYDROLOGICAL MODELING**

by

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ABSTRACT

Large scale landuse changes have been making the news throughout the world. However, the assessment of landuse change impacts on the hydrological cycle is still a challenging task. Complex hydrological models cannot be applied to watersheds without detailed climate, vegetation, soil, and runoff data. Simple models do not provide sufficient support for spatially distributed landuse management decisions. Therefore, this study presents parsimonious, process-based, spatially-distributed hydrologic models to assess effects of landuse changes on runoff in ungauged basins. The introduced models were based upon the assumption that storm runoff is predominantly generated on certain areas of a watershed. The most commonly used method to predict the runoff generation areas is the concept of dominant runoff generation processes (DRP), which are channel interception, subsurface storm flow, Hortonian and saturation excess overland flow.

In particular, forecasts of saturation overland flow generating areas have been controversial in previous research. Therefore, traditional and new soil saturation prediction concepts were evaluated with field data in the first part of this study. Best predictions were found for combinations of topographical indices and groundwater table depths. For three out of four assessed watersheds, optimized model parameters depended on climate. In the second part of the study, DRP model structures were developed. Established DRP area delineation was extended with dynamic process and connectivity modules. The latter were found to improve model fit and parameter feasibility, particularly in process-based DRP models. Temporal connectivity distributions demonstrated that Hortonian overland flow was more affected by connectivity than storm subsurface flow. Third, the DRP models were used to predict stream flow volume, dynamics, and effects of landuse changes on peak flow. Stream flow predictions improved if DRP concepts were added to topographical data; however, the additional DRP-based prediction improvement was marginal if climate data were available. DRP-based snowmelt and peak flow predictions agreed well with observed data. The model was used to predict effects of the mountain pine beetle infestation in British Columbia on peak flow in 290 watersheds. Peak flow increases up to 70% were forecasted, and a strong relationship between peak flow increase and landuse change affected area proportion of a watershed was found.

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LIST OF ABBREVIATIONS

adj. R^2	adjusted coefficient of determination
AGR	agreement statistic (agreement of two binary grids)
ARE	watershed area (explanatory variable in regression models)
BAR	bare surfaces (explanatory variable in regression models)
BC	province of British Columbia (Canada)
BTM	baseline thematic mapping (landuse GIS data set)
CHA	channel interception
CIR	watershed circularity (explanatory variable in regression models)
CN	Cotton North (study area)
CS	Cotton South (study area)
D1ALL	Q90 of the daily specific runoff decreases within one day
D1MAY	Q90 of the daily specific runoff decreases within one day in May
D1OCT	Q90 of the daily specific runoff decreases within one day in October
D5ALL	Q90 of the daily specific runoff decreases within five days
D5MAY	Q90 of the daily specific runoff decreases within five days in May
D5OCT	Q90 of the daily specific runoff decreases within five days in October
D8	D eight (flow delineation algorithm)
DEM	digital elevation model
DHSVM	distributed hydrology soil and vegetation model
DINF	D infinity (flow delineation algorithm)
DRP	dominant runoff generation processes
DP	deep percolation
E	Upper Elk Creek watershed
EIS	event integrated saturation
ELE	elevation (explanatory variable in regression models)
EN	Elk North (study area)
ES	Elk South (study area)
F_{emp}	Empirical cumulative distribution function
FOR	forest (explanatory variable in regression models)
GENOUD	GENetic Optimization Using Derivatives

GIS	geographic information system
GL	Glory (study area)
GLA	glacier (explanatory variable in regression models)
GTS	gradient to the stream (model parameter)
HOF	Hortonian overland flow
HYD	hydromorphic features index (to map soil saturation footprints)
I1ALL	Q90 of the daily specific runoff increases within one day
I1MAY	Q90 of the daily specific runoff increases within one day in May
I1OCT	Q90 of the daily specific runoff increases within one day in October
I5ALL	Q90 of the daily specific runoff increases within five days
I5MAY	Q90 of the daily specific runoff increases within five days in May
I5OCT	Q90 of the daily specific runoff increases within five days in October
LIS	location integrated saturation
M	Malcolm Knapp Research Forest
MAM	mean absolute error was standardized by the maximum observed runoff
MFD	multiple flow direction (flow delineation algorithm)
MOR	morphology index (to map soil saturation footprints)
MPB	mountain pine beetle (MPB, <i>Dendroctonus ponderosae</i> Hopk.)
NDP	no dominant runoff generation process
NSE	Nash-Sutcliffe efficiency
O	Oyster Creek watershed
P	Upper Penticton Creek watershed
PAN	annual precipitation (explanatory variable in regression models)
PLA	wetness indicator plants index (to map soil saturation footprints)
PMA	May precipitation (explanatory variable in regression models)
POC	October precipitation (explanatory variable in regression models)
Q90	90% quantile
QALL	Q90 of the daily specific runoff
QMAY	Q90 of the daily specific runoff in May
QOCT	Q90 of the daily specific runoff in October
R	programming language for statistical computing (www.r-project.org)
R ²	coefficient of determination
RAI	summer rain storms

RC	runoff coefficient
RMSE	root mean squared error
ROS	rain on snow
RSAGA	R package of SAGA
SAGA	System for Automated Geoscientific Analyses
SME	snow melt
SLO	slope (explanatory variable in regression models)
SOF	saturation excess overland flow
SOS	soil and slope index (to map soil saturation footprints)
SSF	shallow subsurface flow
SWE	snow water equivalent
TO	topographical index based on local slope (terrain characteristic)
TOF	topographical index based on the gradient along the flow path (terrain characteristic)
TOPO	topographical index based on local slope (model parameter; e.g. TOPO = 12)
TOPOF	topographical index based on the gradient along the flow path (model parameter)
TSA	two timber supply area
TRIM	Terrain Resource Information Management Program
UNF	unforced runoff generation (runoff without input due to rain or snowmelt)
VE	vertical distance to the groundwater (terrain characteristic)
VERT	vertical distance to the groundwater (model parameter, e.g. VERT = 2m)
VETO	combination of VE and TO
VETOF	combination of VE and TOF

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CO-AUTHORSHIP STATEMENT

Klemens Rosin was responsible for the identification and design of the research program. In general, he performed research and data analyses independently, and wrote the manuscripts of this thesis. The following co-authors contributed to different manuscripts:

- Markus Weiler (University of Freiburg) reviewed data analyses and manuscripts of all chapters.
- Russell Smith (UBC) assisted the field work campaign and modeling study of chapter 2, and reviewed the manuscript of this chapter.
- Georg Jost (UBC) was involved in the study design and data collection of chapter 3.

1 INTRODUCTION

1.1 Global Change Challenges

The world has experienced substantial effects of human induced climate change recently, such as increased average air and ocean temperatures (IPCC, 2007a). Until 2050, further greenhouse gas emission increases and subsequent temperature rises have been predicted (IPCC, 2007b), which could result in considerable consequences for ecosystems, industry, economy, and society (IPCC, 2007c). The relevant process interactions are complex and their effects considerably lagged, possibly beyond one human generation (Underdal, 2010). A variety of global change implications – from economics (Stern, 2006) to migration (Warner, 2010) – are related to water cycle alteration (Bates et al., 2008; Hamlet and Lettenmaier, 2007). The water cycle is expected to change in different ways: For example, lower snow pack (Barnett et al., 2008) could modify snow melt dominated stream flow dynamics (Stewart, 2009). Or extreme precipitation events might occur at higher frequency (Kim, 2005). The hydrologic cycle also plays an important role as a link between climate and landuse change: Leuzinger (2010) reported *direct* impacts of carbon dioxide and temperature increases to be less important for tree growth as their *indirect* role via the hydrologic cycle. On the other hand could climate-induced landuse changes result in considerable alterations of hydrologic processes such as interception, evapotranspiration, infiltration, and percolation (Bonan, 2008).

Nowadays, there is a fairly high agreement among climate change effect predictions on hydrological processes both at national (e.g. Sushama et al., 2006) and regional scale (Rodenhuis et al., 2008). Rodenhuis et al. (2008) reported the following projections for British Columbia until 2050: average temperatures are projected to increase by 1.5°C in winter and summer; average annual precipitation is expected to increase by 6%, with the highest increase in winter and even small decrease in summer; snow pack is expected to decline up to 55% with highest decrease at the coast; glacier volumes will probably be substantially lower; soil moisture is expected to be lower in summer. In contrast to climate predictions, there are no comprehensive predictions available for stream flow for the entire province of British Columbia.

From near future change predictions, new challenges for hydrologic science have emerged (Wagener et al., 2010):

- **Calibration and evaluation:** Stationarity has been questioned (Milly et al., 2008); that is why hydrologic predictions should not always be based on past events and processes. If hydrologic models are calibrated or evaluated with historical data, the effects of ignoring non-stationarity should be carefully assessed.
- **Dominating processes:** Instead of using arbitrary model structures, hydrologic models should carefully represent the dominating hydrologic processes of the assessed watersheds (Sivapalan et al., 2003).
- **Scales:** Runoff generation mechanisms have to be identified at hillslope or watershed scale. However, predictions should be made at relevant scales for society and economy, which are usually substantially larger (Wagener et al., 2010). Hydrological models have to deal with this scale discrepancy.
- **Threshold and cumulative effects:** Process and change are often driven by threshold effects. Furthermore, cumulative effects from different sub-watersheds could result in superimposition or counterbalance (Eaton et al., 2010) and should be considered in a hydrologic model.
- **Spatially-distributed patterns:** Altered processes such as snow melt and evapotranspiration have distinct spatial patterns, which should be captured by a hydrologic model (Croke and Jakeman, 2001).
- **Landuse-hydrology interaction:** Landuse affects hydrologic processes and vice versa; the landuse-hydrology interaction depends on its spatial distribution.
- **Data sources:** Funding for monitoring and measuring sources is expected to decline. However, more digital data at better resolutions will be available (e.g. GRACE; Rodell et al., 2009). Hydrologists should not ignore this trend, and be aware of the value of additional data sources for their predictions.
- **Cross-disciplinarity:** Cross-disciplinary knowledge should be used for reliable hydrological predictions which are transferable to various disciplines.

Future hydrologic models should acknowledge these requirements toward robust predictions of future, human-influenced hydrologic systems. That is why this study presents a new, parsimonious model framework how to predict the effects of landuse changes on stream flow in a changing environment.

1.1 Previous Research

1.1.1 Modeling Landuse Change Effects on Stream Flow

Large scale landuse changes have frequently been reported in the past decade (Lambin et al. 2001, Buergi et al., 2004). Besides direct social, economical, and ecological aspects of land cover changes (Foley et al., 2005), the effects on hydrologic processes have been of interest (Matheussen et al., 2000). In general, it has been accepted that landuse changes could lead to runoff increase (Bosch and Hewlett, 1982). In particular, the influence of logging (e.g. Costa et al., 2003) or climate change (e.g. Middelkoop et al., 2004) on stream flow have been studied in detail. However, most publications focused on effects of landuse change at watershed scale (Fohrer et al., 2001). Only few investigations were made at large scales from a few hundred kilometers to entire continents (e.g. Dudgeon, 2000).

Effects of land use change on stream flow have either been quantified with simple empirical relationships between land cover and hydrologic variables (Leopold, 1968) or fairly complex hydrologic rainfall-runoff models (Storck et al., 1998; Schnorbus et al. 2010). Both methods exhibit substantial disadvantages for applications at large-scale. Simple empirical relations have been developed from paired catchment experiments; however, the experimental results depended on local watershed characteristics, and resulted in substantial variability of predicted landuse change effects on hydrology (Brown et al., 2005). Furthermore, variability due to climate, soil, or geology impedes transference of empirical landuse effects from small to large scale. In contrast to empirical relations, differences of climate, soil, or geology could be considered in spatially, explicit hydrologic models. These models are often physically-based conceptualizations of the hydrologic cycle (Arnold et al., 1998; Wigmosta et al., 1994; Schulla, 1997). For data-enriched watersheds, these models achieved satisfactory predictions of landuse change effects on stream flow (e.g. Bowling et al., 2000). However, complex distributed models are fairly time-consuming at large scale, and rely on extensive watershed and stream flow data for calibration (Van Shaar et al., 2002). Recent projects have consolidated strengths of various complex model structures with ensemble predictions of land use change on stream flow (Breuer et al., 2009), which could not inhibit inherent calibration challenges of distributed models in ungauged basins. Furthermore, the recent tendency toward model selection from the illustrious florilegium of complex hydrologic models has been criticized; the fact that even the smallest models are fairly complex has been described to the point by the title of Goodwin and Wagener (2006) as ‘tall, grande or

venti – appropriate levels of complexity for hydrologic models’. Consequently, new model structures are in demand which allow predictions of landuse change effects on stream flow in ungauged basins. Following Niehoff et al. (2002), successful landuse change effect predictions required definitions of spatially explicit landuse scenarios combined with spatially distributed, process-based hydrologic models. In agreement with these requirements, a new modeling framework was introduced in this study which could be applied to data-scarce (e.g. no soil data) and ungauged basins.

1.1.2 Dominant Runoff Generation Process Models

The lack of calibration data in ungauged basins and corresponding parameter over fitting is a key problem of distributed models. Consequently, there has been a spate of interest in parsimonious, process based models. With high intensity sprinkling experiments and subsequent modeling, Faeh et al. (1997) postulated that one process dominated runoff generation at a certain locations in a watershed. Scherrer and Naef (2003) developed a decision scheme to determine dominant runoff generation processes (DRP). At the same time, Peschke et al. (1999) proposed a method for regionalizing runoff generation processes. In recent years, various rule-based approaches have been introduced to delineate dominant runoff generation processes (Wetzel, 2003; Kirnbauer et al., 2005; Moran et al., 2005; Hellebrand et al., 2007; Markart et al., 2004; Geitner et al., 2009). Most of the new approaches could not be compared directly, since they highly depended on locally available data (Meissl et al., 2008). Consequently, a widely applicable process-based approach should primarily depend on easily available data such as digital elevation models and channel network maps.

1.2 Motivation

The background motivation of this study was to assess effects of the mountain pine beetle (MPB, *Dendroctonus ponderosae* Hopk.) infestation on stream flow in British Columbia (BC, Canada). In the last decade, the MPB has severely decimated lodgepole pine (*Pinus contorta* var. *latifolia*; Taylor et al. 2006). Since 1999, more than 130,000 km² of forest have been infested and 78% of lodgepole pine was expected to be killed until the year 2018 (Kurz et al., 2008). In response to the MPB infestation, the forest industry was salvaging as much timber as possible before it became unusable. Both infestation and logging might have substantial impacts in hydrologic processes such as interception, evapotranspiration, snowmelt, or infiltration. Consequently, the mountain pine beetle could substantially affect peak flow. However, there were only a few

climate and hydrometric stations available in the most affected regions in British Columbia. Furthermore, no soil data sets existed for the entire province of BC.

1.3 Objectives

The objective of this study was to develop a parsimonious, process-based hydrologic model, which could be used to assess effects of landuse changes on stream flow. Existing knowledge about DRP area delineation should be incorporated in the new model; but the model should be extended to dynamic process predictions and interaction among DRP areas. The most critical aspects of the new model should be evaluated with field studies (e.g. soil sampling), observed runoff data, or former research studies. It was planned to test the benefit of DRP concepts for stream flow prediction for different degrees of information available (e.g. topography, landuse, and climate data). The model should be applied to estimate peak flow changes due to the mountain pine beetle infestation in British Columbia.

1.4 Thesis Structure

The model set up of this study followed a hierarchical structure. First, criteria for delineating runoff generation areas were defined and tested. Second, dynamic process models were added to DRP areas. Third, the model was used to predict catchment response and landuse change effects. The following paragraphs describe the most important aspects and the structure of this study.

In general, DRP area delineations have been well established (e.g. Scherrer and Naef, 2003). However, little attention has been given to the role of saturation excess overland flow in temperate forests. Furthermore, several topography-based concepts have been introduced to predict saturated areas (e.g. Ambroise et al., 1996; Merot et al., 2003), but their applicability has not been tested in variable climate regimes as are found in British Columbia. Consequently, chapter 2 of this study focuses on the assessment of soil saturation in forests, and the climatic dependency of parsimonious soil saturation predictions.

In chapter 3, DRP models are defined and evaluated. A comprehensive DRP model consisted of the delineation of contributing areas, dynamic processes, and connectivity of DRP areas and the channel network. The area definitions are based upon previous research and findings of chapter 2. However, only a few studies characterized dynamic DRP models (Hellebrand and van den Bos, 2008). Therefore, different process models were defined and tested for diverse

watersheds and storm types. Even though connectivity of contributing areas has often been discussed (Ali and Roy, 2009), it has never been incorporated in dominant runoff generation process models. Consequently, two simple connectivity concepts were incorporated in DRP models and critically evaluated in chapter 3.

The model structures developed and evaluated in chapters 2 and 3 were used to predict stream flow (chapter 4) and landuse change effects (chapter 5). The stream flow predictions are based on the following ideas: If large proportions of a watershed are assigned to DRP areas, a high runoff ratio could be expected; and if fast runoff generation processes dominated a catchment, flashy response dynamics should dominate the hydrograph. These ideas were used in chapter 4 to evaluate the benefit of DRP concepts for catchment response predictions. The value of DRP concepts might depend on which data sources could be available in addition, which has not been investigated before. Therefore, chapter 4 focuses on DRP-based improvement of stream flow predictions (volume and dynamics) given topographic, landuse or climate data. Chapter 5 presents a framework for applying the DRP concept to predict effects of landuse changes on peak flow.

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2 EVALUATING SOIL SATURATION MODELS IN FORESTS IN DIFFERENT CLIMATES¹

2.1 Introduction

Substantial amounts of stream flow can be generated on areas with water saturated soil (Dunne and Black, 1970). Not only water quantity, but also water chemistry and quality, could be affected by runoff from saturated areas (Molenat et al., 2002). Cappus (1960) distinguished between different runoff generation processes such as overland, near surface, and subsurface storm flow. Betson (1964) observed that runoff was only generated on certain areas in a watershed. Hewlett and Hibbert (1967) described the contributing areas in a watershed to vary during and among storms which led to their development of the variable source area concept. Runoff generating areas have frequently been located in valley floors and on slopes with shapes that are conducive to overland flow generation (Amerman, 1965). It has been demonstrated that often only a small part of a watershed contributes to storm runoff (Ragan, 1968).

Relationships between runoff generation areas and topography have been reported by Kirkby (1975). Beven and Kirkby (1979) used the relevance of topography for runoff generation in the implementation of TOPMODEL - a parametrically efficient model approach relating upslope area, slope gradient and saturated hydraulic conductivity. Due to the availability of low-cost digital elevation models (DEM) combined with increasing computer processing power and the popularity of distributed model problems (e.g. effects of landuse changes), a variety of TOPMODEL approaches has been applied over the past thirty years (Beven et al., 1995). It has been applied not only to predict surface but also subsurface flow (Robson et al., 1992). Topography-based approaches have been critically appraised (Beven, 1997) and often modified: for example, transmissivity (soil-topographic index; Beven et al., 1984), effective rainfall (climato-topographic index; Merot et al., 2003) or the stream network (Mourier et al., 2008) have been included. Grabs et al. (2009) demonstrated that a topographic wetness index derived from a dynamic distributed model represented saturation patterns substantially better than conventional topographic indices.

¹ A version of this chapter will be submitted for publication. Rosin, K., Smith, R., Weiler, M., 2010. Evaluating soil saturation models in forests in different climates.

The contributions of saturation overland flow (SOF) to runoff in forests have mainly been studied in tropical rainforests (e.g. Elsenbeer and Lack, 1996); less attention has been given to the relevance of SOF in temperate forests. Dewalle et al. (1988) found in an oak dominated mixed hardwood forest a mathematical function to relate saturated area growth and contraction to discharge. McGlynn et al. (1999) assessed near-stream saturated area contributions to storm rainfall in a watershed entirely forested with mixed northern hardwoods. However, the coupling of hillslopes to the channel network in temperate forests has often been dominated by preferential flow (e.g. Buttle and McDonald, 2002; Buttle et al. 2004) and not saturation overland flow. Studies about overland flow in temperate forests have often focused on erosion (e.g. Heede, 1987). Governing processes for overland flow and subsequent erosion have been quantified by Miyata et al. (2009) using small-scale hillslope experiments. However, soil saturation areas in temperate forests have rarely been delineated at watershed scale (Blazkova et al., 2002), which is probably more relevant to predict SOF contributions to stream flow than soil saturation analyses at hillslope scale. In watersheds without forest, soil saturation has been efficiently detected at large scale with remote sensing (Mohanty and Skaggs, 2001). However, remote sensing has not been reliably used to measure soil saturation in extremely dense forests (Kite and Pietroniro, 1996). Therefore, soil saturation under forests has mainly been mapped based on soil and vegetation characteristics (e.g. Guentner et al., 2004; Birkel et al., 2010). Most mapping criteria were based on conceptual understanding, and have been evaluated indirectly (e.g. Boorman et al., 1995). However, the soil, vegetation, and morphology mapping criteria have not been directly evaluated with saturation measurements. Furthermore, only a few soil saturation mapping and modeling studies have assessed study areas with clearly inhomogeneous climatic conditions (Merot et al., 2003), and the derivation of model parameter uncertainty (Franks et al., 1998) has not been assessed for different climatic conditions.

Based on these research gaps, this study evaluates simple, mainly topography-based soil saturation prediction models. To test a potential climate-dependence of parameters in topographic models, the evaluation is done in four watersheds with different climate. Due to this large scale, soil saturation is not measured but mapped. Consequently, we will answer the following research questions: I) Which indicators can be used to map saturation, and how can these mapping criteria be evaluated? II) Which indices are the most appropriate to predict soil saturation? III) Does thresholds of topography-based indices depend on the climate? What level

of uncertainty has *a priori* to be taken into consideration when topographic indices are applied to watersheds in different climates?

2.2 Study Sites

Field work was carried out at four locations in British Columbia (BC) (Figure 2.1). Because the climate in British Columbia varies substantially from west to east, the study sites were selected with different distances to the Pacific Coast. Saturation was detected in the Malcolm Knapp Research Forest (M in Figure 2.1); whereas, saturated areas were mapped in Oyster Creek watershed (O), Malcolm Knapp Research Forest (M), Upper Penticton Creek watershed (P), and Upper Elk Creek watershed (E). Table 2.1 presents characteristics of the four study sites. Generally, they could be categorized into two wet (O and M) and two dry (P and E) locations. The two wet sites were located close to the coast at low elevation. In contrast, the dry sites were at high elevation in the interior of British Columbia. All study sites were densely forested.



Figure 2.1: Study sites. The study sites were located in the Province of British Columbia in Canada. O = Oyster Creek watershed, M = Malcolm Knapp Research Forest, P = Upper Penticton Creek watershed, and E = Upper Elk Creek watershed.

Table 2.1: Site characteristics. Properties of the four watersheds Oyster Creek d (O), Malcolm Knapp Research Forest (M), Upper Penticton Creek (P), and Upper Elk Creek (E).

	O	M	P	E
Ecoregion	Eastern Vancouver Island ¹	Lower Mainland ¹	Thompson-Okanagan Plateau ¹	Southern Rocky Mountain Trench ¹
Mean annual precipitation [mm/y]	1387 ²	2194 ³	705 ⁴	580 ⁵
Average temperatures January/August [°C]	2.4/17.0 ²	2.3/17.3 ³	-6.5/12.2 ⁴	-5.8/17.6 ⁵
Elevation range [m]	250-430 ⁶	170-310 ⁶	1605-2020 ⁶	1440-1940 ⁶

2.3 Methods

The objective of this study was to assess soil saturation model structures, which are simple indices. The model structure assessment was based on soil saturation mapping data (Figure 2.2, bottom). The model structures were evaluated in four watersheds with different climate, since potential climate-dependences of the model parameters (i.e. the thresholds of the indices) were of particular interest in this study. But first, the soil saturation mapping indicators were evaluated with simple, effective saturation detection samplers (Figure 2.2, top). Consequently, soil saturation was detected, mapped, and modeled in this study. The spatial resolution of detection and mapping was adapted to the modeling part. For the modeling section of this study, a 25 m digital elevation model (DEM) has been generated for the entire province of British Columbia (Rosin and Weiler, 2008), since the available raw elevation data did not allow generating DEMs at higher resolution.

¹ Environment Canada (2008a)

² Environment Canada, (2008b) for ‘Oyster River UBC’

³ Environment Canada (2008a) for ‘Haney UBC’

⁴ Ministry of Forests, BC (2005)

⁵ Jost et al., 2007

⁶ Terrain Resource Information Management Program (GeoBC, 2007)

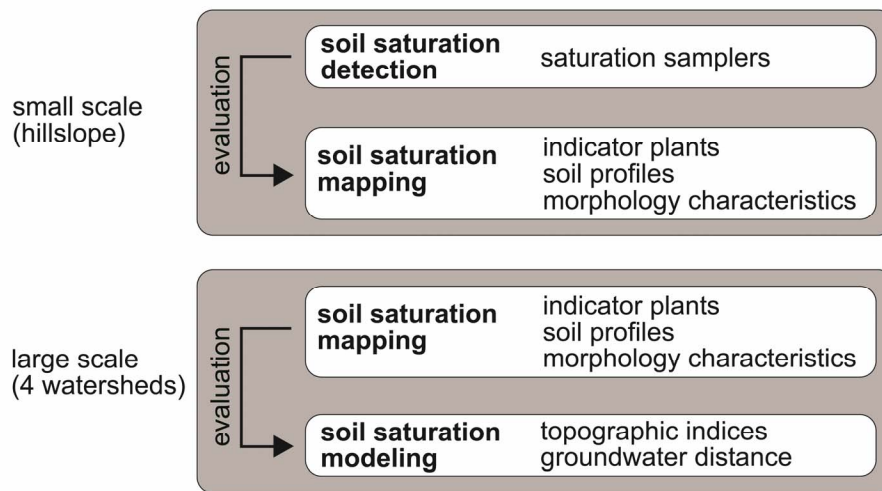


Figure 2.2: Method overview. Soil saturation mapping was used to evaluate the model structures (i.e. indices). The mapping was evaluated with detected soil saturation.

2.3.1 Saturation Detection

To detect soil saturation, simple saturation samplers were developed and tested. They consisted of PVC bottles with holes for water-inflow and air-outflow (Figure 2.3, right). The hole for water-inflow was positioned one centimeter below the soil surface so that the sampler would fill with water upon saturation of the soil surface. Data were collected in Malcolm Knapp Research Forest from May to December 2007. In this period, 70 samplers were checked twelve times after rainstorms varying between 6 and 96 mm/day. The samplers were installed in fourteen randomly selected 25 m grid cells in a forested sub-watershed (Figure 2.3, left). Estimation of saturation characteristics utilized five randomly located samplers per grid cell. Due to considerable workload of installing and checking the samplers, they could not be applied to detect saturation over entire watersheds. Consequently, field procedures were developed to efficiently map saturation patterns for larger watersheds.

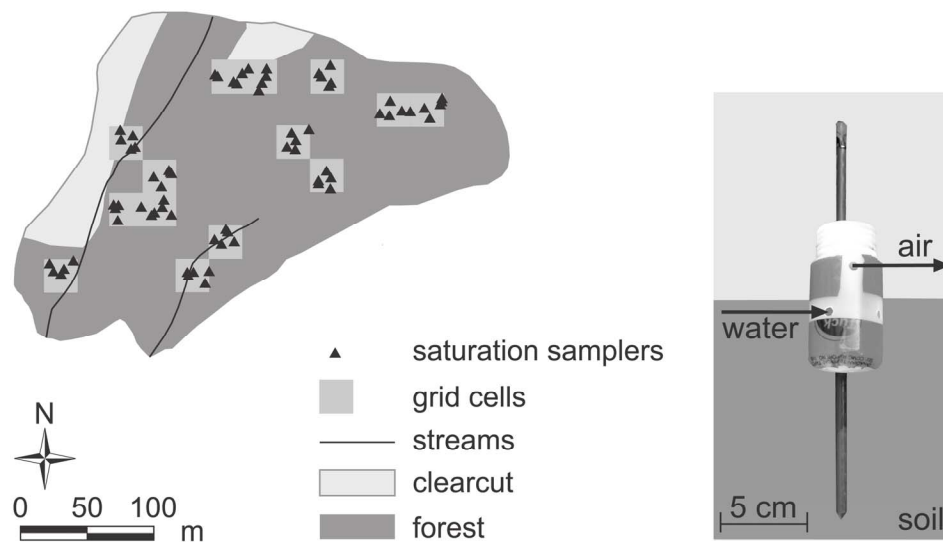


Figure 2.3: Location and design of saturation samplers. 70 saturation samplers were installed at random positions within each randomly selected 25 m grid cell (left). A stake fixed the plastic bottles at the soil surface (right).

2.3.2 Saturation Mapping

Saturation mapping was based on time-integrated soil saturation measures, as applied by Guentner et al. (2004). This approach incorporated long-term saturation patterns, which integrated soil saturation over seasons and years, and, therefore, should not depend on the timing of the assessment. Our study distinguished between primary and secondary saturation indicators. Primary indicators were not or only little climate-specific (e.g. wetness indicator plants and hydromorphic features). For example, the presence of wetness indicator plants indicated wet conditions over long periods. However, secondary indicators were clearly related to the climate (e.g. soil texture, morphology) and, thus, only represented relative saturation for a given climate. For example, soil saturation could be found on a silt-clay but not on a sandy-silt in a wet climate. However, in a dry climate not even the silt-clay soil might be saturated. In this example, silt-clay soil could be an indicator of saturation in a wet climate, but not in a dry climate. Since soil saturation depends on multiple factors including the *combination* of climate and soil texture, the latter might explain some saturation differences within one climate but not across climates. Because of ambiguity associated with the secondary indicators, this climate-orientated study was mainly based on primary saturation indicators. However, it was possible to underestimate saturation with primary saturation indicators (e.g. indicator plant scarcity due to light shortage). To address this issue, secondary saturation indicators were used to detect underestimation of

saturation within each watershed. To estimate primary and secondary saturation indicators, characteristics from indicator plants, soil, morphology, and micro-topography were mapped (Table 2.2).

Table 2.2: Mapping characteristics. Characteristics of soil, plants, and morphology that were mapped to describe the propensity of soil surface saturation.

Class	Measures
Indicator plants	Total abundance of plants
	Abundance of wet indicator plants
	Abundance of moist indicator plants
Soil	Hydromorphic features
	Humus depth
	Soil depth
	Texture
Morphology	Longitudinal slopes
	Latitudinal morphology
	Longitudinal morphology
	Slope position
Micro-topography	Roughness height

Field data were collected between July and October 2007. A transect mapping approach was selected with approximately half of the transects starting at the stream network. For each transect, average characteristics of entire 25 m grid cells were mapped. The mapped qualitative information was translated into indices which expressed the propensity of soil saturation. Because the time-integrated measures could also be perceived as footprints of saturation, the final mapping indices were called footprint indices. In this study, four footprint indices were defined: plant (PLA, primary indicator), hydromorphic features (HYD, primary indicator), soil-slope (SOS, secondary indicator), and morphology (MOR, secondary indicator).

2.3.2.1 Plant Index

Each grid cell was mapped with respect to plants indicating frequent soil saturation. We focused on plants without considering trees and mosses, since this study assessed topsoil saturation. The mapping was not based on the leaf area, but on plant abundance, in order not to overestimate large-leaved indicator plants. According to Klinka (1989), Pojar (1994a), and Pojar (1994b), the indicator plants were classified into two categories, MOIST and WET. WET plants included the following species: *Blechnum spicant*, *Lysichitum americanum*, *Rubus spectabilis*, *Streptopus roseus* and *Trillium ovatum*. MOIST plants included *Polystichum munitum* and *Dryopteris expansa*. The final wetness plant index (PLA) was defined to equal one for 100% abundance of WET plants. We subjectively assumed that 100% MOIST plant abundance corresponded to

similar wetness conditions as if 30% WET plants were observed. The combination of WET and MOIST plant abundance to a single wetness plant index *PLA* was expressed as

$$PLA = 0.003 \cdot M + (0.01 - 0.00003 \cdot M) \cdot W \quad (\text{Eq. 2.1})$$

where *PLA* is the plant index, *W* the percentage of WET, *M* the percentage of MOIST plant abundance per grid cell. Abundance of WET plants in the entire grid cell resulted in a *PLA*-index of 1.0. Similarly, the plant index equaled 0.3 for 100% MOIST and no WET plants. Equation (2.1) is displayed in Figure 2.4.

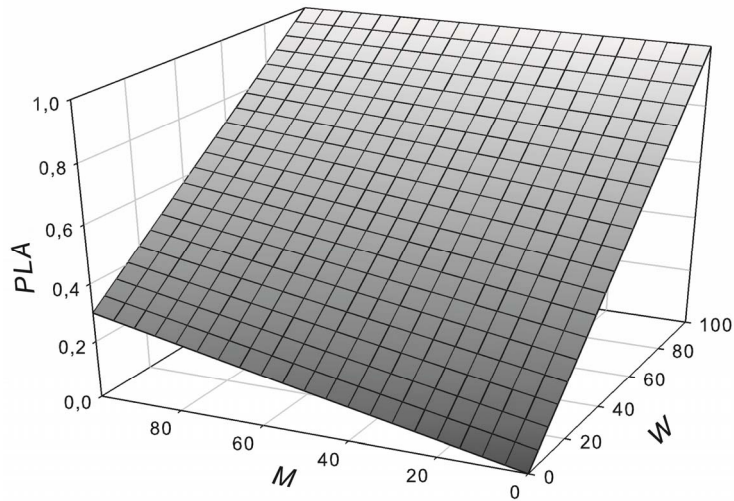


Figure 2.4: Wetness indicator plant index (*PLA*). *M* = MOIST plants [%], *W* = WET plants [%].

2.3.2.2 Hydromorphic Features Index

In order to estimate soil properties, two soil profiles were examined in each 25 m grid cell: one in the wetter and the other in the drier portions of the cell. For apparently homogeneous grid cells, only one soil profile was analyzed. If hydromorphic features were found in the top 30 cm of all soil profiles in a grid cell, the hydromorphic features index (HYD) was set to 1.0. If two profiles were assessed, but only one contained hydromorphic features, HYD was set to 0.5. If neither of the profiles contained hydromorphic features, HYD was set to zero. Since the HYD index could only take three values, it was applied in combination with the plant index.

2.3.2.3 Soil-Slope Index

The Ministry of Forests of British Columbia developed a guide for site identification based on a biogeoclimatic ecosystem classification (Ministry of Forests, BC, 1994). It incorporated a key to assess the relative soil moisture regime with environmental features, which were used in our study to translate mapped information into the soil-slope index (SOS). This index was based on soil texture, soil depth, slope gradient, slope position, and the presence of hydromorphic characteristics. Information from all soil profiles in each cell was averaged; if qualitative information (e.g. soil texture) differed between the two soil profiles, the more representative information for a given grid cell was selected based on additional soil samples from hand auger holes.

2.3.2.4 Morphology Index

Small scale surface morphology characteristics and micro-topography could store water even in steep slopes. To accommodate morphology-influenced soil saturation, we defined a morphology index MOR:

$$MOR = b + 0.6 \cdot \max \left(0, 1 - \frac{S}{20 + \frac{60 - 20}{10} \cdot R} \right) \quad (\text{Eq. 2.2})$$

where S is slope [%], R roughness height [m] (see Figure 2.5 for a definition of R), and b a concavity parameter. The parameter b was defined according to the concavity along ('longitudinal') and perpendicular ('latitudinal') to the fall line (Table 2.3). During our field studies, we visually observed that concavity dominated the soil saturation: in general, concave sites exhibited high and convex sites low saturation. Due to the high relevance of concavity, the concavity parameter b in Equation 2.2 defines a preliminary range of the MOR index. MOR fell between 0.4 and 1 for concave (Figure 2.5, center), and between 0 and 0.6 for convex sites (Figure 2.5, right panel).

Table 2.3: Site concavity parameter. The parameter b described the site concavity and ranged from 0 to 0.4. Categories (concave, plain, convex) were defined based on a threshold slope difference of five degrees (from the grid cell center to two points up- and down-slope at a distance of 12.5 m).

parameter b		latitudinal to the fall line		
		concave	plain	convex
longitudinal to the fall line	concave	0.4	0.3	0.2
	plain	0.3	0.2	0.1
	convex	0.2	0.1	0.0

Among sites of similar concavity, soil saturation has been observed to vary depending on microtopography and slope (Loos and Elsenbeer, 2009). We described microtopography of a grid cell with its roughness height R , which was defined as the deviation of micro-hills from a theoretical plane whose slope and elevation equaled the average slope and elevation of the grid cell (Figure 2.5, left). The observed roughness heights mainly fell within 0 to 7 m. However, we assumed that the interaction of slope and microtopography controlled soil saturation at given concavity. It was assumed that on slopes S steeper than 60%, not even extreme roughness heights of 10 m could keep water in the hillslope (Figure 2.5, center and right). On the other hand, we assumed that on slopes less than 20% no microtopography was needed to cause soil saturation (Figure 2.5, center and right). The combination of concavity, slope, and roughness was quantified with Equation 2.2. Figure 2.5 contains MOR surfaces according to Equation 2.2 for $b = 0.4$ (center) and $b = 0$ (right).

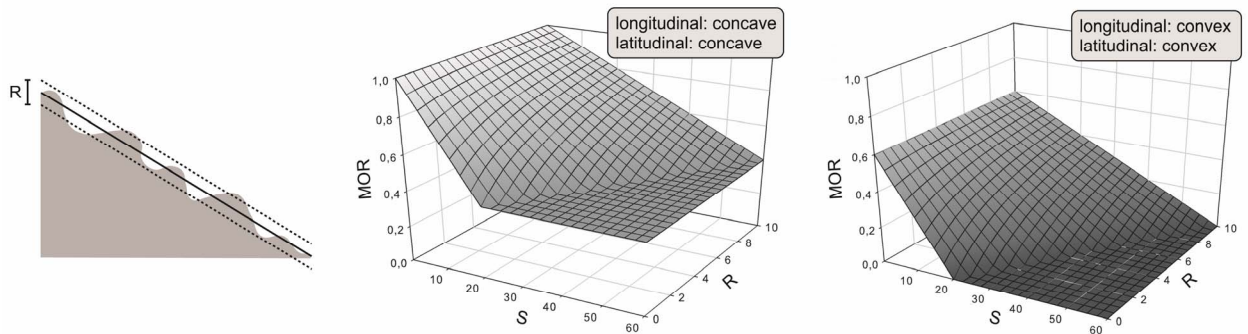


Figure 2.5: Morphology index (MOR). Left: Definition of the roughness height R . Center: MOR for a concave grid cell (both longitudinal and latitudinal to the fall line). Right: MOR for a convex grid cell. R = roughness height [m], S = slope [%].

2.3.3 Modeling Saturation

In this study, topography-based soil surface saturation indices were evaluated. Despite considerable simplification, our topography-based indices attempted to address the relevant processes. With respect to modeling saturated areas, saturation could either occur from upslope overland flow or a rising groundwater table. Saturation patterns have often been found in riparian areas resulting from the latter process (McGlynn and McDonnell, 2003). Therefore, the vertical distance to the stream (VE) was assessed as a potential index to predict saturation. Saturation could also occur in localized flat areas with large upslope areas, which resulted in a high topographic index (TO), defined as $\ln(A/\tan(\alpha))$ where A is the upslope area per unit contour length and α is the local slope gradient (Beven and Kirkby, 1979). Similarly, Merot et al. (1995) proposed a topographic index where the slope represented the gradient to the stream along the flow path (TOF), which took into consideration the drainage capability from a certain location to the stream network. Hjerdt et al. (2004) proposed that a down slope topographic wetness index represented groundwater gradients more realistically, and demonstrated that this index has a lower dependence on both DEM resolution and local slope.

The topographic index is not defined for slopes equal to zero, yet flat areas could often be important for soil saturation. In order to eliminate grid cells with a slope of zero, three solutions were proposed and tested (Figure 2.6): addition of a constant values all grid cells (a), replacement of all zeros by a constant value (b), or superposition of non-constant values (c). The problem of slope equal zero also arose when the gradient to the stream was selected as the relevant slope (Merot et al., 1995). The procedures (b) and (c) caused inconsistent shifts within a DEM. Consequently, procedure (a) was applied in this study. The selected value of an additional constant was central for our approach — if a small constant was added, the difference to the original slope was marginal; however, this caused high topographic index values independent of the upslope area. For example, a constant of 0.01 degrees resulted in a topographic index of 15.1 for an upslope area of one 25 m grid cell. On the other hand, adding a large constant value changed the slopes of the entire grid considerably. As a compromise, a value of two degrees was added to all slopes in this study.

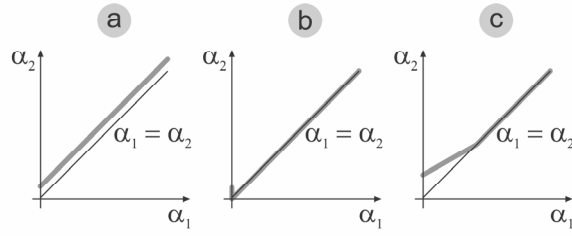


Figure 2.6: Slope alteration in flat areas. Possible procedures are shown, how to address slopes equal zero (α_1 = original slope, α_2 = slope applied in the index): Addition of a constant value to all grid cells (a), replacement of zeros (b), or superimposition of non-constant values (c).

In addition to the three indices VE, TO, and TOF, we predicted soil saturation with two combinations of them (VETO, VETOF). For the combined indices a grid cell was considered saturated if saturation was predicted either due to the topographic index (TO/TOF) or due to the groundwater criterion (VE). The combined indices were incorporated because saturation could be induced from water flowing down slope while perched on less permeable layers as well as due to a groundwater table being close to the soil surface, particularly in locations close to water bodies.

In this study, the calculation of the topographic index generally followed the approach by Beven and Kirkby (1979). The soil-topographic index (Moore et al., 1986) was not considered, because transmissivity data were not available for the watersheds. In RSAGA (Brenning, 2008), the vertical distance to the water table (VE) was estimated from the digital elevation model and the channel network assuming a homogeneous and stationary aquifer in time and space (Olaya, 2004). The depth of the water table was interpolated from the stream network; the shape of the water table estimated from the curvature of surface according to the digital elevation model (Olaya, 2004). The water bodies had distance of zero to the groundwater, and provided basic elevation values for the VE interpolation. Channel cells were defined from 1:20,000 topographic maps, which were provided for the entire Province of British Columbia by the Terrain Resource Information Management Program (TRIM). The TRIM digital elevation model (25m by 25m cells) was also used in this study. To improve the existing TRIM DEM, the original point elevation data (xyz) and breaklines were included to generate a new seamless digital elevation models for the entire Province of British Columbia (Rosin and Weiler, 2008). The spatial analysis of this study was coded with the programming language R, particularly with the package RSAGA (Brenning, 2008), which was based on SAGA 2.0.2 (SAGA, 2008).

2.3.4 Comparing Observed and Predicted Saturation

2.3.4.1 Small Scale Analysis

The fourteen grid cells in Malcolm Knapp Research Forest, where soil surface saturation samplers were installed (Figure 2.4), underwent saturation mapping in order to determine the four footprint indices (Section 2.3.2). This allowed comparison of the mapping indices with saturation detection (Section 2.3.1). The Malcolm Knapp area was chosen because it consisted of both very dry and very wet portions, similar to soils in different climates in British Columbia. The occurrence of saturation and the time-integrated saturation measures were compared using linear regression. Furthermore, detected saturation was graphically compared to precipitation.

2.3.4.2 Large Scale Analysis

Simulated soil surface saturation was evaluated at large scale in four watersheds with different climates. For this evaluation at least 100 grid cells were mapped in each watershed. The sample size per watershed was derived from a theoretical sampling study. The bases of the sampling study were potential saturation locations in the four watersheds, which were derived from preliminary field studies in combination with topographic index estimations. For each sample size the saturation proportion in a watershed was calculated for 100,000 different theoretical sampling locations. In all watersheds the sampling variability of saturation proportions decreased substantially for samples sizes greater 70. Consequently, 100 grid cells were mapped in each watershed. The mapping data of over 400 grid cells across four study areas were used to evaluate indices and to estimate parameters (i.e. thresholds of the indices) and uncertainty ranges.

In this study, appropriate indices were determined based on the agreement of saturation according to footprint indices and predicted soil surface saturation. Grid cells in a watershed could be classified into cells with agreement and disagreement between saturation footprints and indices (Table 2.4).

Table 2.4: Mapping and index agreement. Grid cells are categorized according to the agreement between mapped footprint saturation (first subscript) and modeled saturation indices (second subscript). The amount of cells in a watershed with agreement equals n_{++} plus n_{--} ; the amount of disagreement cells in a watershed is the sum of n_{+-} and n_{-+} .

		index	
		saturated (+)	not saturated (-)
mapping	saturated (+)	n_{++}	n_{+-}
	not saturated (-)	n_{-+}	n_{--}

We evaluated indices based on the agreement of mapping and saturation index (AGR), which is defined as

$$AGR = \frac{n_{++} + n_{--}}{n} \quad (\text{Eq. 2.3})$$

where n_{++} is the amount of grid cells in a watershed which are saturated according to mapping and indices, n_{--} is the amount of grid cells which are not saturated following both mapping and indices, and n is the total amount of grid cells in a watershed. AGR corresponds to the ‘observed agreement’ by Cohen (1960).

For the index evaluation, each assessed grid cell of the four watersheds should be assigned to be ‘saturated’ or ‘not saturated’. However, there were two difficulties to overcome toward a binary saturation classification of the mapping information: 1) Different footprint indices (e.g. PLA, HYD) contain diverse saturation information. 2) There is no clear threshold for each footprint index above which saturation could be expected. First, we assumed that if either the PLA-index or the HYD-index indicates saturation at a certain location, this grid cell is considered to be saturated. Second, we did not determine a single threshold, but multiple probability weighted thresholds. High probability was assigned to thresholds for which we expected (based on our saturation sampling study) saturation above and no saturation below this value. For the PLA index, 1000 thresholds following a symmetric triangular probability distribution from 0.15 to 0.25 (maximum = 0.2) were chosen. For HYD, probabilities of 0.7 (for HYD = 0.5) and 0.3 (for HYD = 1) were selected. The total agreement between footprint saturation (for each combination of PLA and HYD) and saturation index consisted of weighted agreements for different PLA and HYD thresholds.

An uncertainty range was defined as the maximum agreement minus 5%. A smaller value (e.g. 0.5%) would only describe the numeric inhomogeneity of the local maximum; a larger value (e.g. 20%) would also incorporate not acceptable parameter values. We believed that the maximum minus 5% was a trade-off that represented the uncertainty of optimum model parameters. Figure 2.7 illustrates both the maximum agreement and the uncertainty range. The AGR curves consisted of several steps: if, for example, the TO-threshold was increased, AGR might decrease because some cells with footprint-saturation (which were modeled correctly) were no longer simulated as saturated. However, after an additional increase of the TO-threshold, dry-footprint-cells (which were modeled incorrectly) were simulated as not-saturated. Consequently, several maxima might occur in the AGR-curve. Favorable model indices should exhibit high agreement and high parameter identifiability represented by a narrow uncertainty range.

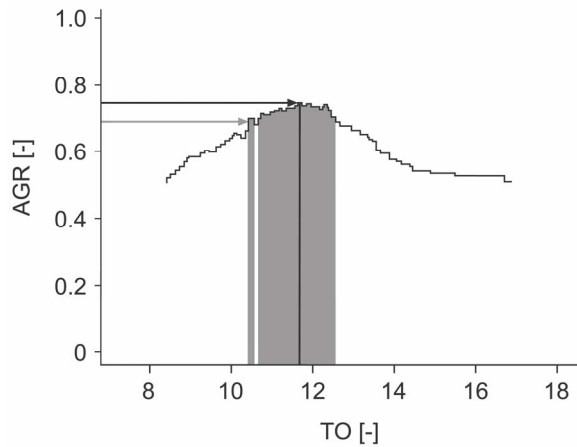


Figure 2.7: Agreement between footprint saturation and index. As an example, the agreement (AGR) is shown for different parameters of the topographic index model structure (TO) for the Oyster Creek watershed. Black arrow and line = maximum agreement; grey arrow and area = maximum agreement minus 5%.

2.4 Results

2.4.1 Small Scale Analysis

During large fall rainstorms, we visually observed distinct saturation patterns in the sub-watershed within the Malcolm Knapp Research Forest. Event Integrated Saturation (EIS) was determined individually for each cell as the percentage of filled samplers per grid cell averaged over all studied events from May to December 2007. First, the percentage of filled samplers was computed for each event; the average over all events resulted in EIS. Generally, the highest EIS values were found close to streams and exceeded 20% (Figure 2.8); however, EIS values of more than 10% also occurred for grid cells further away from the stream.

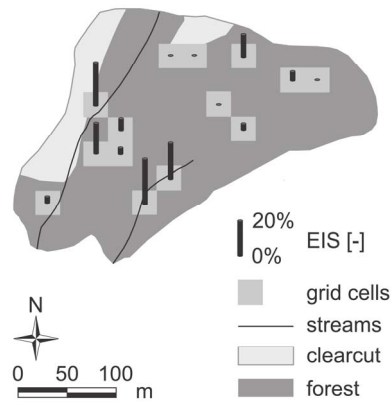


Figure 2.8: Event integrated saturation. Event integrated saturation (EIS) based on detection with saturation samplers is shown for a sub-watershed in the Malcolm Knapp Research Forest.

The four footprint indices were compared to EIS for the fourteen sampled grid cells (Figure 2.6). In general, the higher footprint indices were found at locations where more samplers indicated saturation. This was true for all footprint indices, even though the coefficients of determination (R^2) varied from 0.26 (HYD), 0.81 (SOS), 0.84 (MOR) to 0.90 (PLA). Furthermore, the PLA values were substantially lower than SOS and MOR across all EIS levels. No linear regression was estimated for HYD in Figure 2.9, since hydromorphic features could only be found in one grid cell. This corresponded to the study by Thompson (1994) at a nearby site, in which a lack of hydromorphic features even at sites with frequent saturation was reported.

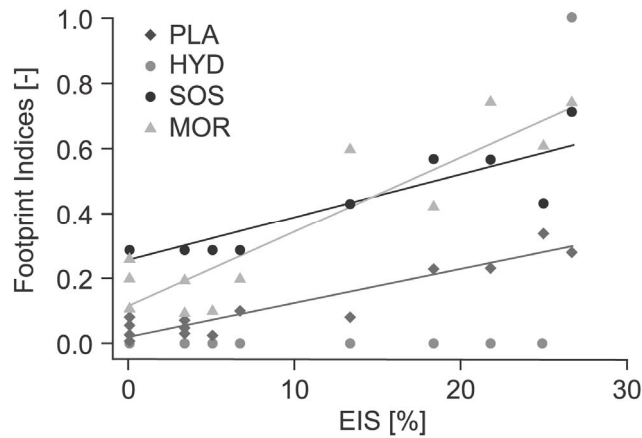


Figure 2.9: Footprint indices and event integrated saturation. Values for the four footprint indices were compared to event integrated saturation (EIS) at fourteen grid cells in the Malcolm Knapp research forest. EIS is the percentage of filled saturation samplers per grid cell integrated over twelve observed storms. The colors of the regression lines correspond to the point colors.

Location Integrated Saturation (LIS) was determined individually for each event as the percentage of filled samplers averaged over all cells, and was used to relate the observed saturation levels to precipitation rates. Figure 2.10 displays LIS values and the maximum daily precipitation in periods between the examinations of the saturation samplers. In summer, no relationship could be observed. However, in fall there was a positive asymptotic relationship with the curve leveling off for large storms with return periods of approximately two years. The maximum area saturated for this sub-watershed was approximately 20%.

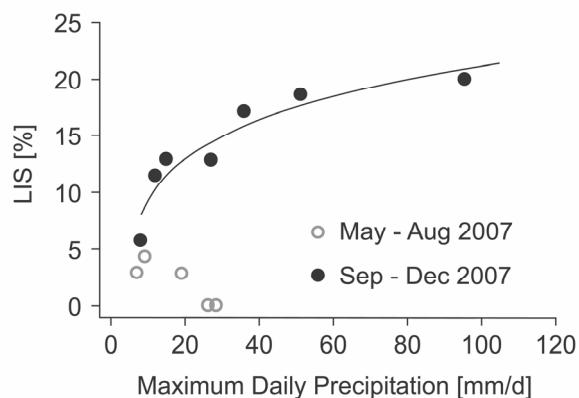


Figure 2.10: Location integrated saturation and precipitation. Location integrated saturation (LIS) was compared to the maximum daily precipitation for twelve periods in 2007 between saturation sampler examinations.

2.4.2 Large Scale Analysis

The objective of the large scale analysis was to select feasible indices for predicting spatial saturation patterns, to estimate the index thresholds, and to assess possible climate dependencies. The index evaluation was predominantly based on the PLA and HYD footprint indices. The secondary footprint indices (SOS, MOR) were only used to identify grid cells where saturation was possibly underestimated with PLA and HYD.

Potential underestimation of the primary footprint indices (e.g. plant scarcity due to light shortage; see Section 3.2) was assessed with secondary indices SOS and MOR. Forty-eight grid cells were classified as low saturation according to the primary saturation mapping indices ($PLA \leq 0.3$ and $HYD = 0$), but high saturation according to the secondary saturation mapping indices (SOS or $MOR \geq 0.6$). Thresholds were subjectively selected to distinguish, definitively, between wet versus dry conditions.

The raw mapping data of these 48 grid cells were analyzed in detail. In three grid cells, the total abundance of plants (not only wetness indicator plants) ranged from 5 to 40%, yet the high secondary footprint indices suggested sufficient water supply. It was unclear whether the low abundance of plants was due to a shortage of water, light or nutrients. Because of this discrepancy, the three grid cells were not included in further analysis. The omitted grid cells were all located in Malcolm Knapp Research Forest, which is covered by the densest forest of all four study locations.

The agreement of the five indices with the primary footprint indices was assessed at large scale with distances of more than 400 km among the four studied watersheds. Figure 2.11 displays the maximum agreement between footprint and indices for the four assessed watersheds. In accordance with theoretical reasoning, the agreement with the combined indices (VETO/VETOF) was at least as high as agreement with TO, TOF or VE. The additional contribution from combining indices was small in the Upper Elk Creek watershed (E); but AGR was higher for combined indices in the Oyster Creek watershed (O) and Malcolm Knapp Research Forest (M). Depending on how model complexity was emphasized, either simple or combined indices could be preferred. Among the simple indices, topographic index with the slope along the flow path (TOF) generally agreed best with footprint saturation. VE had smaller agreement than other indices for the O and E watersheds. Similar to the simple indices, VETOF

exhibited a slightly higher agreement than VETO. The topographic index with the gradient along the flow pathway performed better than the topographic index using the local slope.

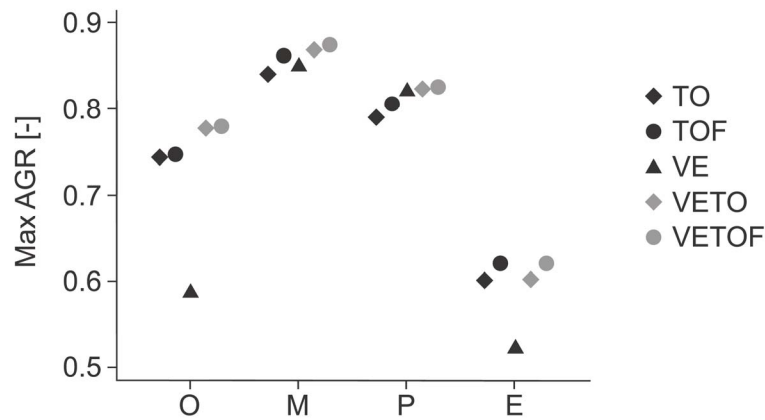


Figure 2.11: Maximum agreement between modeled indices and footprint indices. The maximum Agreement (Max AGR) of different indices with footprint saturation is displayed for the four assessed watersheds (‘wet’: O and M; ‘dry’: P and E).

The differences in agreement were larger among watersheds than among indices; however, no consistent differences between wet (O and M) and dry watersheds (P and E) could be observed. High agreements were found across all indices in M and P. Agreement in E was lower than for the other watersheds. The parameter values that showed the highest agreement between the different modeled indices and the footprint saturation were assessed using not only the maximum agreement, but also parameter uncertainty. The results are displayed in Figure 2.12. The four graphs on the left side are based on TO indices, while the four graphs on the right side refer to TOF indices. Each row represents the analysis of one watershed. The results of the *single* index evaluation are shown with bars along the x-axis (TO or TOF) or the y-axis (VE). The squares display the results of the *combined* index evaluation. For consistency, the results of the VE single index are shown in all graphs.

For all watersheds, the differences in agreement between TO and TOF were marginal due to their high correlation. The coefficients of correlation between TO and TOF were 0.96 (O), 0.97 (M), 0.98 (P), and 0.98 (E). The parameter values with highest agreement were similar for VETO and VETOF.

For watersheds O, M and P, the best parameter values and corresponding uncertainty ranges seemed to exhibit climate dependent patterns for all five indices. The optimum parameter values for TO and TOF were lower and the optimum for VE was higher for the two wet watersheds (O and M) than for the dry watershed (P). For the E-watershed, the optimum value for VE was low as well; however, the optimum values for TO and TOF were lower than expected for this dry watershed. The same patterns could be observed for simple (TO, TOF, VE) and combined indices (VETO, VETOF). The agreement showed distinct maxima for the O and E watersheds but wide maxima and, therefore, higher parameter uncertainty for the M and P watersheds.

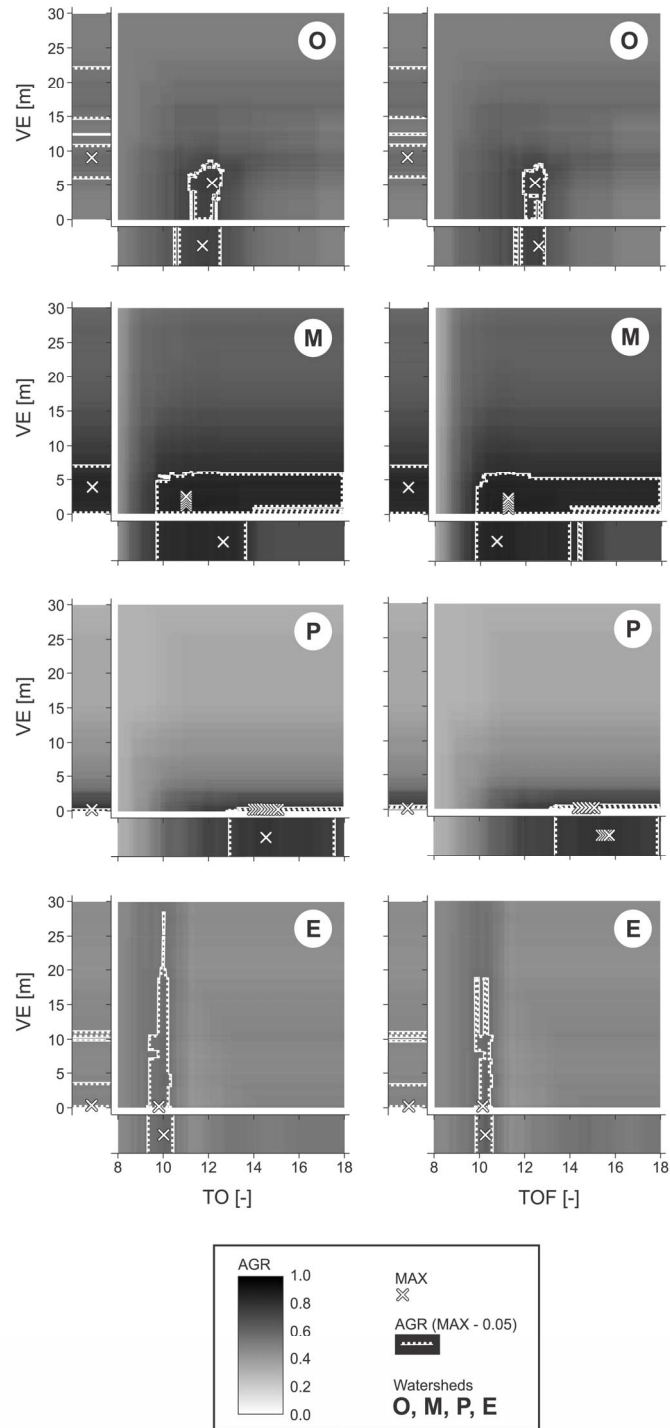


Figure 2.12: Agreement between model and footprint indices. The agreement (AGR = grey scale) between model and footprint indices is displayed for the four watersheds (rows) and five indices (TO, TOF, VE, VETO, VETOF). Left column: TO and VE indices, right column: TOF and VE indices. Furthermore, the maximum agreement (MAX = white crosses) and the maximum agreement minus 5% (white crenelated lines) are shown.

2.5 Discussion

2.5.1 Small Scale Analysis

Four aspects of the small scale analysis for the Malcolm Knapp Research Forest were particularly relevant for the subsequent evaluation of saturation mapping and modeling procedures. First, the small scale analysis underpinned the hypothesis that saturation could frequently occur on forested watersheds in humid climates. Second, the saturation patterns helped to understand the processes that cause saturation, which helped with identifying the processes that had to be represented by the model. Third, the selected mapping criteria could be evaluated with the saturation measurements, which were the most important findings of the small scale analysis. Fourth, the comparison of saturation to precipitation established a relationship from the saturation based approach - from saturation detection, to mapping, to modeling a hydrologic system. These four aspects are discussed in sections 2.5.1.1 to 2.5.1.4.

2.5.1.1 Soil Surface Saturation in Forests

The relevance of soil saturation and SOF in tropical rain forests is beyond dispute (Elsenbeer, 2001). Loos and Elsenbeer (2009) reported a median spatial overland flow generation frequency of 0.421 (maximum = 0.926), which was larger than our median LIS values of 0.086 (maximum = 0.200). The higher saturation occurrence in tropical forests seemed to occur mainly due to the higher frequency of intensive storms.

In temperate forests, SOF has often been neglected since shallow subsurface flow has been reported as the most important runoff generation process (Peters et al., 1995). Near soil surface runoff has often been detected and assigned as flow over organic layers or shallow lateral flow rather than saturation overland flow (Buttle and Turcotte, 1999). In forested watersheds in the northern hemisphere, the importance of SOF has been noted for minor storms; whereas, for wet conditions, subsurface flow has been found to be the major source for runoff generation (Sidle et al., 2000). Even though our analysis only concentrated on surface saturation and not on subsurface flow, the amount of event integrated saturation (EIS) values above 20% was considerable and supported the relevance of saturation in temperate forests. The high saturation levels in our study (Figure 2.5) suggested that SOF might contribute considerably to runoff generation. The visually observed saturation patterns in Malcolm Knapp Research Forest exhibited spatial extents of 5 to 10 m. Hence, five randomly distributed samplers per 25 m grid

cell were considered sufficient to capture the observed spatial variability within each grid cell while still being manageable in terms of required workload. However, EIS probably overestimated the temporal persistence of saturation over the entire period, since saturation was only measured after intensive storms. The event integrated saturation was only useful as a qualitative measure to compare the propensity of saturation among grid cells.

2.5.1.2 Saturation Generating Processes

Frequency and location of detected saturation led to preliminary interpretations about processes contributing to saturation. On the western part of the sub-watershed, a distinct channel was located in a steep valley. Due to the short hillslopes, lateral contributions from upslope areas to grid cells next to the channel were marginal, yet saturation samplers close to the stream were frequently saturated. We assumed that saturation was, to a certain extent, caused by the *rising water tables*. For minor storms in the summer, only water table-triggered saturation was observed. That is in agreement with Moore and Thompson (1996) who observed the highest water tables in convergence zones as ephemeral channels and lower slopes. Our observations also agreed with the findings by McGlynn and McDonnell (2003) who found that runoff from riparian areas as a fraction of total runoff was considerable for small events. On the other hand, some of the upslope located saturation samplers in the eastern part of the watershed were filled with water after intensive storms, particularly at locations with *moderate slopes* within generally steep contributing areas. For these upslope locations, water table contributions to saturation were unlikely. However, saturation only occurred at locations with moderate slopes within generally steep contributing areas. We assumed that, at these sites, saturation occurred due to the accumulation of water flowing down slope. Waldenmeyer (2003) has reported saturation patterns in similar locations according to an ecological moisture index. In summary, locations and timing of saturation indicated that two different processes generated soil saturation, which were referred to as riparian and hillslope saturation by McGlynn and McDonnell (2003). We define riparian saturation as soil saturation to be induced by fluctuating water tables; on the other hand, we define hillslope saturation to be predominantly controlled by topography-based upslope flow.

2.5.1.3 Mapping Criteria Evaluation

The small scale analysis was also used to evaluate the suitability of the footprint indices. Saturation occurrence has been mapped in previous studies (e.g. Guentner et al., 2004), but little attention has been given to evaluating the mapping criteria. Even though three footprint indices

(PLA, SOS and MOR) predicted saturation reasonably well, a few implications from the footprint indices required discussion. First, the index resolution was critical for the assessment. Particularly for the SOS and HYD index, coarse classifications might reduce the coefficients of determination in Figure 2.6. Second, it was expected that no index would perfectly represent saturation for all locations. To get an idea of the variability across footprint indices, we assessed the three footprint indices separately. However, in future studies, a subjective likelihood measure could be introduced to rate the adequacy of a footprint index for a specific location. For example, if the indicator plants did not seem to provide distinct saturation information for a certain location, a low likelihood could be taken into account. In addition, a higher spatial and temporal density of saturation observations would be beneficial for future saturation model evaluations. More frequent measurement (e.g. 15 minute intervals) could allow comprehensive inference about the duration of saturation. This would allow for an elaborate relationship of saturation and precipitation that would be more informative than only monitoring intense precipitation (e.g. Scherrer and Naef, 2003). Automatic approaches (e.g. Srinivasan et al., 2000) combined with wireless sensor technology (Trubilowicz et al., 2009) would be valuable for future studies.

The integration of the mapped field data with the footprint indices required parameterization (e.g. Equation 2.1). The integration parameters could be calibrated against measured saturation, but calibration would only be justified for a high density of saturation measurements. In this study, the integration procedures were based on subjective process understanding. The fundamental idea was to define time-integrated, season-independent measures of saturation footprints. Even though this idea was predominantly realized with the definition of the four footprint indices, a few violations occurred. For example, the depth to the water table was part of the soil-slope index, which depended on the meteorological conditions before the field assessment. The outcome of the indicator plant analysis was probably not completely independent of the season; however, the dependency was reduced by assessing the abundance of plants rather than the leaf area.

2.5.1.4 Saturation and Precipitation

The location integrated saturation (LIS) was used to estimate the contributing area. For larger storms, we observed saturation in more than 15% of the watershed, which was substantially higher than the contributing area of 1.2 to 3% reported by Ragan (1968) in a forested watershed. However, the maximum daily precipitation of 33 mm in Ragan (1968) was considerably lower

than maximum daily precipitation of 96 mm in our study. On the other hand, the saturated area in our watershed was comparable to values observed by Dickinson and Whiteley (1970), where storms of 90 mm were related to a theoretical minimum contributing area of 20%. Waldenmeyer (2003) estimated that saturation overland flow was generated in 15% of his forested watershed. Our LIS value is in correspondence with Hutchinson and Moore (2000) who estimated surface contributing areas as 428 m² (~21%) during a ten day winter storm in a nearby study site.

2.5.2 Large Scale Analysis

2.5.2.1 Mapping Strategies

For the large scale analysis, different mapping strategies were considered such as mapping of entire watersheds (e.g. Schmocker-Fackel, 2004), mapping of randomly selected single cells, transect mapping at random locations in the watershed, and transect mapping from the stream network (Blazkova et al., 2002). The approach by Schmocker-Fackel (2004) was too labor-intensive for large, forested watersheds with few roads. The random single cell approach was not appropriate to comprehensively capture transitions from the wet riparian to possibly dry upslope zones. In order to get a distinct idea of the wetness transitions, the selected method involved mapping transects of 50 to 150 m perpendicular to the stream, which was similar to the approach introduced by Blazkova et al. (2002). Unlike Blazkova et al. (2002), we also surveyed transects that did not intersect any streams in order to capture unexpected saturation patterns on seemingly dry slopes. Furthermore, we mapped entire 25m grid cells to derive average characterizations per grid cell. With 100 grid cells mapped per watershed, the comprehensive saturation mapping could be managed for grid sizes of 25 m. Since topographic data were not available at higher resolution, the study was based on 25 m grids. Nevertheless, we were aware of substantial shortcomings. For example, Thompson and Moore (1996) found reduced predictive power of topographic indices on observed water table levels as the grid resolution was decreased from 4 to 8 to 16 m, since runoff generation dominating topography features were at lengths less 4 m. In particular, effects of slope curvature and microtopography are not represented by low resolution grids (Moore and Thompson, 1996). With a simulation study, Zhang and Montgomery (1994) reported limited process representation power of a 25 m grid; they suggested that a 10 m DEM could be a reasonable trade-off between flow representation and computational demand. However, it has also been reported that predictions with parsimonious models do not necessarily improve with increasing grid resolution (Watson et al., 1998). For example, terrain analysis with

a 30 m digital elevation model could be used to estimate average water table depths at this grid resolution, since at this prediction errors might average out (Thompson and Moore, 1996).

2.5.2.2 Model Structures

In this section, the two basic indices (the topographic index and the vertical distance to the groundwater table) are discussed. Apart from known limitations of the topographic index (see e.g. Beven, 1997), two practical implications are reviewed: the estimation algorithms for the local slope and the estimation algorithms for the upslope area.

A variety of algorithms exist for calculating the local slope, which essentially differ in how many adjacent cells are included in the calculation. The maximal slope (e.g. Travis et al., 1975) has been described not to provide realistic flow patterns and some optimizations have exhibited limitations with respect to their mathematical robustness (Costa-Cabral and Burges, 1994). However, second- or third-order polynomials have represented slopes fairly realistically (Zevenbergen and Thorne, 1987; Haralick, 1983). We evaluated the most common slope calculation methods by assessing the modeled slopes with the surveyed slope gradients of more than 400 grid cells in the four watersheds. Based on this assessment, slopes calculated with third-order polynomials were much too steep. Consequently, second-order polynomials were used in this study to estimate local slopes. This has been described to be the most prominent method (Olaya, 2004), which made our approach comparable to other studies.

Different methods have been proposed to calculate upslope areas; they were highly related to slope estimation concepts. Unidimensional flow algorithms have not represented realistic flow patterns (e.g. D8; O'Callaghan and Mark, 1984) and some bi-dimensional flow algorithms (e.g. MFD) have produced considerable dispersion. This did not allow for a unique upslope area definition. In contrast, D-infinity has been shown to minimize dispersion and generate realistic flow patterns (Tarboton, 1997), and was used in this study to estimate the upslope area. In general, we are confident that the saturation dynamics of the shallow soils in British Columbia are consistent with the assumptions of a topography based model. This is supported by Thompson and Moore (1996) who found statistically significant relations between water-table depths and topographic indices in a nearby study site.

Not only the topographic index model structures, but also vertical distance to the groundwater (VE) contained substantial methodological uncertainty. In contrast to the depth-to-water index by Murphy et al. (2009), VE was an approximation to derive the vertical distance of a grid cell to a theoretical groundwater table (Etzrodt et al., 2002). It was estimated with RSAGA (Brenning, 2008), which required digital layers of the channel network and digital elevation models. In RSAGA, the elevations of the channels were calculated first, a possible groundwater table was interpolated with iterations considering realistic hydraulic gradients second, and the distance from the groundwater table to the surface was computed third. With this method, neither soil heterogeneity nor geological properties were included in the estimation of the groundwater table. VE is only meaningful if a detailed and realistic stream network were available. For the Oyster Creek watershed, where no digitally registered streams fall within a close range, the VE index was not capable of reproducing the groundwater table realistically. For a reasonable application of an index based on the vertical distance to the stream (VE, VETO, and VETOF), a stream network density of at least 1 km/km² was proposed.

2.5.2.3 Model Evaluation Measures

For evaluating spatially distributed binary patterns of simulated and observed data, Kappa statistics have often been suggested (e.g. Congalton and Green, 1999). In hydrology, Kappa statistics have frequently been applied for evaluating saturation indices (e.g. Guentner et al., 2004). However, they have been described to be a highly controversial method to compare the agreement of binary patterns (e.g. Feinstein and Cicchetti, 1990). Many paradoxes of Kappa statistics have been assessed, such as the dependence on bias and prevalence (e.g. Hoehler, 2000). Byrt et al. (1993) showed examples where the Kappa was higher for a lower level of agreement. Because the highest Kappa value was assigned to an index without the highest agreement to the field data, we believed that Kappa statistics were not an appropriate method for model evaluation and comparison. Furthermore, a ‘chance-corrected measure’ (Byrt et al., 1993) was not required in our study to determine model agreement with field data.

Alternatively, we applied the proportion of agreement (AGR) between model and field data as a meaningful measure for model evaluation. Since correct delineation of saturated areas is the foundation of many hydrologic models, the model evaluation allowed both cases of false modeling (either model=saturated / field data=not saturated, or model=not saturated / field data=saturated) to equally falsify the runoff contribution area. This was consistent with an equal

treatment of both modeling errors of AGR. Because it was necessary to assess agreement between the model and the field data at the position of each grid cell to ensure spatially correct distributed modeling, a model evaluation based on the average saturation ratio in the watershed was deemed inappropriate. Consequently, we chose AGR instead.

2.5.2.4 Model Structures and Parameters

The best model parameters (i.e. index thresholds) and uncertainty ranges varied considerably within and between watersheds. In order to assess the impact in parameter uncertainty on contributing areas, the estimated parameters and uncertainty ranges were applied to mesoscale basins (20 to 35 km²), within which the four watersheds of the large scale analysis were located. For each mesoscale basin, proportions of saturated areas were calculated according to the best model parameters and their uncertainty ranges. The corresponding distributions are displayed in Figure 2.13. Steps in the uncertainty range resulted in steps in contributing areas (e.g. VE for the O-watershed in Figure 2.11).

In general, the contributing area varied more among basins than among indices. Based on our limited process understanding of the selected mesoscale basins, the saturation proportion of watershed E was relatively high. Furthermore, the ranges of the saturation ratios according to the uncertainty measures AGR (MAX – 0.05) were considerable for all indices and basins. The uncertainty range was small only in watershed P with a small proportion of contributing area. The uncertainty in predicting saturation was not reduced if a combined index like VETO and VETO_F was applied.

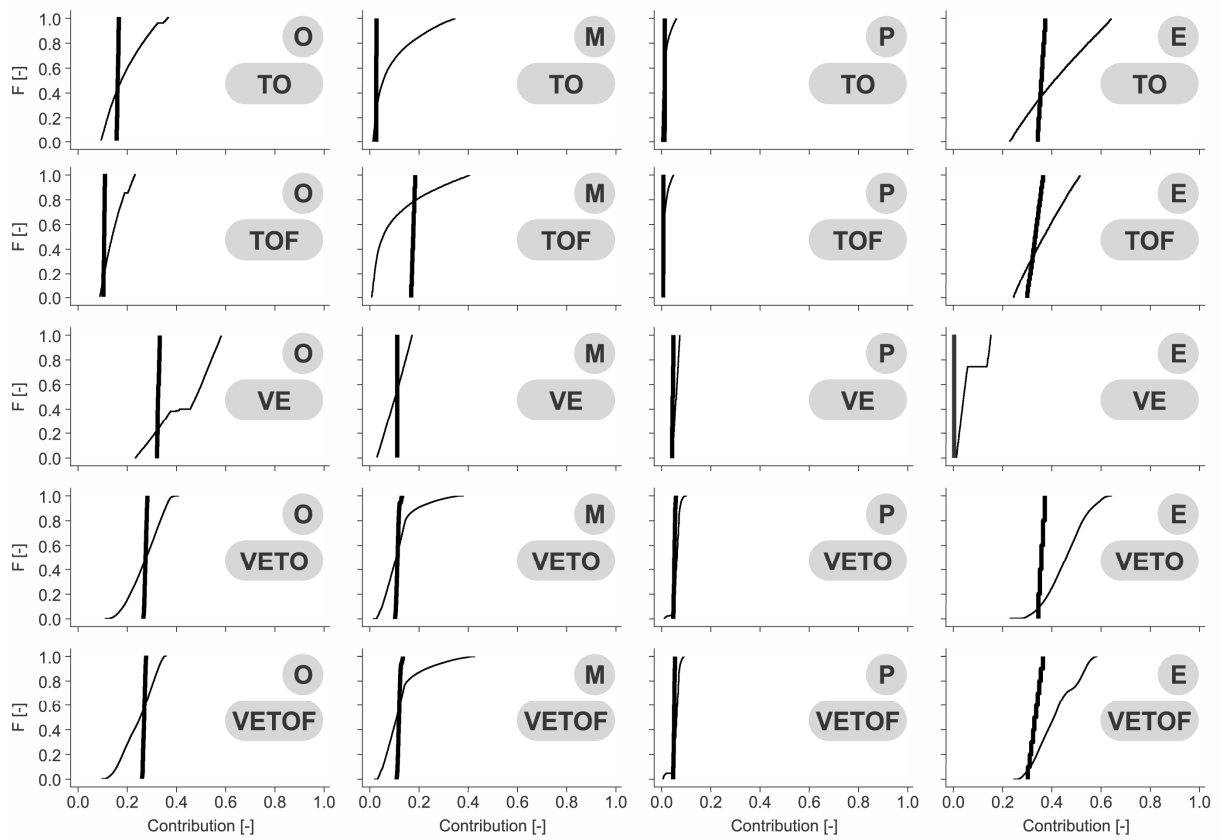


Figure 2.13: Runoff generating watershed proportions. Empirical distributions (F) of the saturation proportions of a watershed ($Contribution$) are shown according to the best model parameters (MAX = thick line) and the uncertainty ranges (AGR (MAX – 0.05) = thin line) for four basins (O, M, P, E) and five indices (TO, TOF, VE, VETO, VETOF).

For the combined indices, the contributions from different model components were assessed (Figure 2.14) and were found to vary substantially for different basins. In general, the agreement of VE was found to be limited (Figure 2.10); however, Figure 2.14 shows that the contribution of VE in VETO/VETOF was not negligible. Consequently, the VE approach could be meaningful, but only in combined indices. For the Oyster Creek basin, the locations of saturated areas for the best index thresholds of different indices are visualized in Figure 2.15.

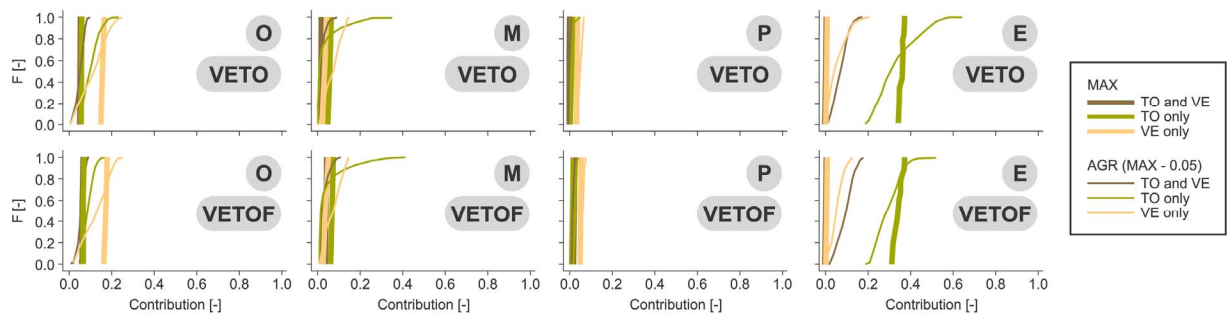


Figure 2.14: Model component contributions. Contributions to runoff generation are shown for different components (TO, TOF, VE) in combined indices (VETO, VETOF) for four basins (O, M, P, E).



Figure 2.15: Saturated area predictions. Predicted saturated areas according to the best model parameters of two indices (TOF, VETOF) are displayed for the Oyster Creek basin.

The contributions of the different indices (Figures 2.14 and 2.15) can be used to infer driving forces and first order controls on soil saturation. In our analysis, we compared the parameters of topography-based indices to precipitation without incorporating topology and typology (Buttle, 2006). Based on our visual field observations, we believed that the specific source of the inflows (i.e. vertical versus lateral sources) might not be as important as the level of down slope connectivity (i.e. rate of lateral outflow) in controlling the occurrence of saturation for the predominantly shallow soils in British Columbia. This speculation is supported by Hutchinson and Moore (2000) who found reasonable relationships between theoretical upslope contributing areas and measured through flow only for lowest flows; however for higher flow, the relationship was worse, probably due to the orientation of dominating active macropores.

Since the parameter variability was notable (variability of optimum parameter sets according to AGR in Figure 2.12), it could be presumed that additional model indices describing other runoff generating processes might improve the estimation of the saturation patterns. On the other hand, parameter estimation uncertainty could increase with model complexity, which might be another potential danger of multiple saturation criteria: an additional criterion would only be meaningful if the supplementary saturated cells contributed to a distinct and realistic saturation pattern. We found only a minor additional agreement by combining indices (Figure 2.8).

In general, more grid cells were assigned as saturated due to the TO than the VE-criterion. However, the ratios of VE- to TO-contributing cells differed substantially for the four watersheds: 0.24 (O), 0.55 (M), 0.51 (P), and 0.22 (E). In this ratio, cells that fulfilled both the TO and VE criterion were included in the numerators and denominators of the calculated ratios. The VE contribution was often considerably higher compared to ratios of riparian to hillslope area contributions of 0.14 reported by McGlynn and Seibert (2003). This difference might be explained by the steep slopes in British Columbia, which was a reason why saturated areas were often located in valley bottoms close to streams. Furthermore, McGlynn and McDonnell (2003) reported a proportional increase in the contribution of hillslope runoff for large events, and a proportional decrease in the contribution of riparian runoff. This might raise the question of whether our long-term saturation based mapping criteria benefited from the detection of saturation close to streams, since these areas were also saturated after small storms. An imbalance of the mapping criteria could also lead to an overestimation of the VE/TO ratio by the model. Further studies should investigate how mapping criteria might depend on precipitation duration and intensity.

2.6 Conclusions

This study provides a framework to assess indices and index thresholds to predict saturation patterns in watersheds in different climates. The proposed approach tries to take into consideration an understanding of the relevant runoff generation processes to delineate meaningful indices. Our research questions are addressed with the following conclusions: I) Wetness indicator plants, hydromorphic features, soil properties and morphology characteristics contained time-integrated saturation information. However, soil and morphology only provide relative, climate-dependent information about saturation. Saturation footprint indices for indicator plants, soil-slope, and morphology, which were derived in this study, correlated

reasonably well with observed soil saturation. II) Saturated areas could potentially be predicted with an index based on a topographic index model combined with modeling the distance to the groundwater table. However, for one of the four studied watersheds, agreement of the saturation model with surveyed soil saturation indicators was much lower than for the other three watersheds. Furthermore, a combined model index (topographic index and vertical distance to groundwater) improves the prediction only slightly. III) For the three watersheds with high agreement between mapping and modeled index, a climatic dependence of the index thresholds was observed. However, no clear relationships between index thresholds and precipitation could be developed for all four watersheds. Therefore, substantial *a-priori* parameter uncertainty should be considered when topography-based indices are applied to watersheds with different climates.

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3 CONNECTIVITY IN DOMINANT RUNOFF GENERATION PROCESS MODELS¹

3.1 Introduction

It is widely accepted that storm runoff is predominantly generated in limited areas of a watershed (Dunne and Black, 1970). These observations were formulated into the partial contributing area concept (Betson, 1964) and extended to the variable source area concept (Hewlett and Hibbert, 1967). This theory has been used to divide catchments into hydrological active and passive zones (Engman and Rogowski, 1974). The spatial distribution and density of contributing areas have been considered to be key controls on runoff formation (Dickinson and Whiteley, 1970).

On the basis of these concepts, Naef and Scherrer (1998) defined dominant runoff generation processes (DRP) for intensive rain storms. Through extensive sprinkling experiments they identified controls on DRP on hillslopes with different soil type and geomorphology. The experimental findings resulted in the creation of a decision scheme to delineate DRP areas (Scherrer and Naef, 2003). These experimentally assessed DRP were simultaneously evaluated with a complex modeling study (Faeh et al., 1997). The iterative process of experiment and model comparison resulted in an improved, more comprehensive understanding of dominant runoff generation controls (Scherrer et al., 2007). Up to now, the DRP concept was predominantly applied in central Europe to assess storm flow generation potentials. For example, DRPs were assessed in alpine (e.g. Markart et al., 2004; Austria), agricultural (e.g. Schmocker-Fackel et al., 2007; Switzerland), and partly forested watersheds (Hellebrand et al., 2007, Casper, 2002; Waldenmeyer, 2003; Germany/Luxembourg). Recently, Meissl et al. (2008) compared four DRP delineation schemes (Scherrer and Naef, 2003; Peschke et al., 1999; Loehmannsroebe et al., 2000; Markart et al., 2004) in an alpine catchment, and found substantial differences in delineated DRP areas among the four studies.

Past research has mainly focused on the delineation of DRP areas (e.g. Meissl et al., 2008) and the description of hydrologic processes within DRP areas (Scherrer et al., 2007). However, little attention has been given to the interaction of water fluxes among DRP areas. Despite its potential

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impact on water quantity and quality no model structures have been formulated which predict the connections between contributing areas in DRP models. It is still unclear if connectivity should be considered in dominant runoff generation process models or other similar modeling frameworks. Yet, it may be highly valuable to know for which storm types or watershed characteristics connectivity should be considered. This may be especially relevant for studies dealing with changes in spatial distribution of DRP areas (e.g. land use change studies). For water quality problems it is also important to know which areas are connected to the channel network and for what duration (e.g. Lane et al., 2004).

The concept of connectivity is well established in ecology to understand interactions between population dynamics and habitat fragmentation (e.g. Andreassen et al., 1998). Recently, connectivity approaches have become more popular in disciplines such as hydrology, geomorphology and sedimentology. Despite frequent applications by different disciplines, connectivity concepts have not been used in a unifying, converging way (Ali and Roy, 2009). Even though the concept of connectivity could play a unifying role between disciplines such as hydrology and ecology (e.g. the connectivity of river systems has impacts on hydrologic processes as well as on population dynamics), interdisciplinary connectivity research is rarely addressed (Tetzlaff et al., 2007). Furthermore, definitions and understanding of connectivity varied substantially between and even within disciplines (Lexartza-Artza and Wainwright, 2009). As a result, connectivity has been applied as a *passe-partout* concept (Ali and Roy, 2009). However, recent literature reviews (Lexartza-Artza and Wainwright, 2009) and classifications (Ali and Roy, 2009) have helped to clarify definitions toward a common understanding of connectivity concepts. In this study, we distinguished between ecological, sedimentological, and hydrological connectivity. We further differentiated hydrological connectivity into riverine connectivity (connections of river systems; e.g. Wiens, 2002) and hillslope - channel network connectivity (e.g. Jencso et al., 2009). Hillslopes can be connected for a short duration (e.g. due to overland flow or shallow subsurface flow) or long duration (e.g. due to deep subsurface flow and groundwater flow). This study only considered the short duration hillslope - channel network connectivity. We defined the hillslope - channel network connectivity as a system state for which water from hydrological active locations of a hillslope contributes to storm runoff in the channel network by overland or shallow subsurface flow.

In reality, it is difficult to identify overland and subsurface flow paths (Knudby and Carrera, 2006). This is why hydrological connectivity has rarely been addressed directly, but quantified with intermediate variables such as pH value (Tetzlaff et al., 2007) or electrical conductivity (Stieglitz et al., 2003). Most publications about hydrological connectivity have relied on conceptual approaches with experimentally quantified indicator variables. An overview of connectivity-related experimental designs has been provided by Ali and Roy (2009). However, little attention has been given to modeling connectivity, in particular to modeling hillslope - channel network connectivity.

A few hydrological connectivity models have been implemented at small scale, for which detailed information about soil properties was available. For example, Deutsch (1998) introduced a simulation code to assess connectivity in three dimensions based on detailed permeability and porosity data. Kirkby and Bracken (2005) assessed long term erosion equilibria with a one dimensional sediment transport model. The potential of connectivity models has been demonstrated with synthetic data sets. Knudby and Carrera (2006) used theoretical fields with different degrees of connectivity for flow and solute transport simulations. They found the hydraulic diffusivity to be an adequate indicator of connectivity. Based on synthetic data, Pardo-Iguzquiza and Dowd (2003) introduced an algorithm how to quantify connectivity with functions and three dimensional statistics. Some studies tried to combine theoretical modeling approaches and experimental knowledge; simulations by Reaney et al. (2007) showed that connectivity increased at lower slopes, which matched experimentally observed high infiltration rates. Hillslope - channel network connectivity has been modeled for fairly small watersheds with detailed soil property information. For example Stieglitz et al. (2003) modeled flow connectivity in watersheds with a modified TOPMODEL version, in which shallow subsurface storm flow simulations considered the spatial distribution of soil depths. Stieglitz et al. (2003) demonstrated that connectivity of upslope locations to streams depended on hydrologic processes on mid slope areas; the mid slope areas were key controls for the connectivity of the entire watershed, which predominantly occurred for snowmelt and storm events with high antecedent soil moisture. Grayson et al. (1992) incorporated connectivity in a simple distributed rainfall-runoff model that was based on topography and soil data; surface roughness and antecedent soil moisture exhibited a substantial effect on the outcome of the simulations. Lane et al. (2004) defined connectivity with a TOPMODEL modification, which estimated minimum topographical indices along flow paths from channels upslope; a location was only considered to contribute to storm runoff if its

down slope areas along a flow path generated runoff as well. The results of connectivity simulations at the watershed scale have been used for decision support in hydrological or ecological questions. Heathwaite et al. (2005) modeled surface, subsurface, ditch, and drain flow including flow connectivity for land use management decisions. Most connectivity models at the watershed scale rely on very detailed soil information. However, little attention has been given to modeling hillslope - channel network connectivity in watersheds with scarce data available. However, new connectivity approaches, which do not rely on detailed soil data, are required for DRP models. The potential of DRP models for predictions in ungauged basins without comprehensive data calls for adequate connectivity modules.

The aim of this study was to identify under which conditions implementing connectivity might improve model predictions or process understanding. Consequently, different approaches for modeling connectivity were tested for a number of DRP models in combination with varying storm types and watersheds. The practical background of this research was the assessment of the hydrologic consequences of the Mountain Pine Beetle infestation in British Columbia (Canada) with a DRP-based rainfall-runoff model (Carver et al., 2009). We were interested to know if and under which circumstances connectivity has to be considered in DRP models. Specifically, we were interested in four aspects of connectivity in DRP simulations: 1) How can the connectivity concept be integrated in DRP area delineation and process models? 2) What are the effects of modeling connectivity on the size and spatial distribution of DRP areas? 3) Do connectivity modules improve DRP models? 4) Are some dominant runoff generations processes more affected by connectivity than others? 5) What are the computational costs of simple connectivity modules?

This study presents a framework for the evaluation of connectivity in process-based hydrological models. We are interested to know under which conditions connectivity needs to be considered in dominant runoff generation process models. The research results could support the design of connectivity modules of spatially distributed hydrological models. In the first part of the methods sections the different model structures are explained, which consisted of the modules DRP area delineation, runoff generation, and connectivity. The second methods section describes the study site, model input data, and analytical methods. The results section deals with effects of considering connectivity on contributing areas, model quality, effects between different runoff

generation processes, and computational costs. The findings of this study – with emphasis on differences between model structures – are critically reviewed in the discussion section.

3.2 Methods

3.2.1 Model Structures

3.2.1.1 Overview

Four dominant storm runoff generation processes (DRP) were considered and simulated in this study: (1) channel interception (i.e. precipitation which falls in streams or lakes; CHA), (2) Hortonian infiltration excess overland flow (HOF), (3) saturation excess overland flow (SOF), and (4) shallow subsurface storm flow (SSF). The developed rainfall-runoff model consisted of three modules: a) DRP area module, b) runoff generation module, and c) connectivity module. The three modules are described in detail in the following sections and Figure 3.1.

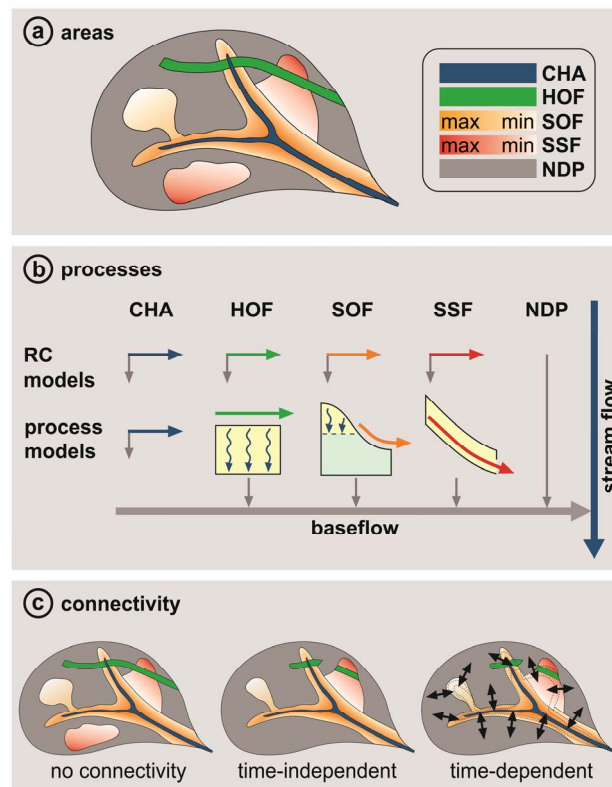


Figure 3.1: Model structures. a) Runoff generation area module b) runoff generation module c) connectivity module. Four dominant processes CHA, HOF, SOF, and SSF were simulated. Areas without dominant runoff generation processes were named NDP (no dominant process). Runoff coefficient (RC) and process models were compared.

3.2.1.2 Runoff Generation Area Delineation

Runoff generation areas were delineated for each dominant process. The spatial analyses of this study were based on a 25 m grid. All grid cells were divided into dominant runoff generation process (DRP) or no dominant runoff generation process (NDP) cells. The runoff generation areas were classified with simple, mainly topography-based rules, since detailed soil survey data were not available (Table 3.1). Three parameters to delineate the DRP areas are described in Table 3.1. Carver et al. (2009) provided a more detailed description of the DRP delineation.

Table 3.1: DRP area definition criteria.

DRP	Area delineation criteria	Parameters
CHA	Streams and lakes.	No parameter
HOF	Forest roads and recently burned areas.	No parameter
SOF	Areas with high topographic indices and/or low vertical distances to the groundwater (Etzrodt et al., 2002).	<i>TOPO</i> : Threshold of the topographic index. <i>VERT</i> : Threshold of the vertical distance to the groundwater.
SSF	Locations with high average gradients to the stream along the flow path.	<i>GTS</i> : Threshold of the average gradient to the stream.

3.2.1.3 Process Hierarchy

Following Scherrer and Naef (2003), one process often dominates the runoff generation at a given location. In our study, one dominating runoff generation process was assumed for each grid cell. Due to the classification criteria in Table 3.1 it was possible that several processes occurred at one location; therefore, processes were hierarchically ordered from faster runoff generation processes to slower processes. SSF was subordinate to SOF, HOF, and CHA. The result of the classification was a DRP map in which each grid cell was assigned to either CHA, HOF, SOF, SSF or no dominant runoff generation process (NDP).

3.2.1.4 Runoff Generation Models

Two different approaches to simulate runoff generation on pre-defined DRP areas were tested in this study: 1) simple runoff coefficient models, and 2) more realistic process models. For each model type runoff generation mechanisms were defined for each dominant runoff generation process, which are described in the next two paragraphs.

3.2.1.5 Runoff Coefficient Models

The model parameters of the runoff coefficient model type were a runoff coefficient (RC) defined as the proportion of runoff to input in each cell for each runoff generation process (RC_CHA, RC_HOF, RC_SOF, RC_SSF). RC_CHA reflected the stream width considering that not entire 25m grid cells acted as streams. In the runoff coefficient models it was assumed that water not contributing to runoff adds to baseflow (Figure 3.1b). For the groundwater storage, a simple linear reservoir was defined. The outflow constant (K_{GW}) of the groundwater reservoir was not optimized, but was estimated *a priori* from the recession of the hydrographs during dry periods.

3.2.1.6 Process Models

The conceptual models of the process based runoff generation mechanisms were defined by knowledge and experience from hillslope experiments (Scherrer et al., 2007; Graham, 2009; Tromp-van Meerveld and Weiler, 2008). Channel interception was simulated with runoff coefficients like in section 3.2.1.5. For HOF, a maximum infiltration rate TH_{HOF} was defined assuming that infiltrating water (below the infiltration rate) contributed to groundwater recharge; infiltration excess water was assumed to run off as overland flow (Figure 3.1b).

The SOF generation depended on the soil saturation deficit which was represented by the maximum storage capacity of a linear reservoir. We assumed that the maximum saturation deficit (MSD) of each SOF-cell depended on an upper storage limit of the reservoir (S_{UP} ; constant over the entire watershed) and location-dependent saturation likelihoods. The location-dependent saturation likelihoods were combinations of the saturation likelihoods due to the local topographic index (L_{TO}) and the local vertical distance to the groundwater (L_{VE}). Equation 3.1 defines the relationship between MSD , S_{UP} , L_{TO} , and L_{VE} . High saturation likelihoods (L_{TO} , L_{VE}) result in small saturation deficits (MSD).

$$MSD = S_{UP} \cdot (1 - \max(L_{TO}, L_{VE})) \quad (\text{Eq. 3.1})$$

High saturation likelihoods were assumed for grid cells with a high topographic index or low vertical distance to groundwater. The saturation likelihoods (L_{TO} , L_{VE}) are defined as

$$L_{TO} = \frac{TO - TOPO}{TO_{max} - TOPO} \quad 0 \leq L_{TO} \leq 1 \quad (\text{Eq. 3.2})$$

$$L_{VE} = \frac{VE - VERT}{VE_{min} - VERT} \quad 0 \leq L_{VE} \leq 1 \quad (\text{Eq. 3.3})$$

where TO is the topographic index and VE the vertical distance to the groundwater of each grid cell, $TOPO$ and $VERT$ parameters of the SOF area delineation (see Table 3.1), TO_{max} the maximum possible topographic index for the study area ($TO_{max} = 18$), and VE_{min} the minimum possible vertical distance to the groundwater ($VE_{min} = 0$). The saturation likelihoods were linearly interpolated between maximum and minimum values of the topographical index respectively the vertical distance to the groundwater (Equations 3.2 and 3.3). For grid cells with no saturation deficit, additional water input resulted in saturation overland flow. An outflow based on a linear reservoir concept was added to groundwater recharge (Figure 3.1).

For SSF, we assumed that subsurface flow within the shallow soils followed the overland flow paths. Therefore, average gradients to the stream along overland flow paths were used to estimate ratios of lateral to vertical flows (L/V), which is defined by the triangle of forces concept as

$$L/V = \sin(\arctan(GS)) \quad (\text{Eq. 3.4})$$

where GS are gradients to stream of different grid cells. For grid cells with gradients to the stream (GS) equal zero only vertical flow occurs; for gradients of 90 degrees the L/V -ratio equals one. Due to varying soil and bedrock permeability the real ratio of lateral to vertical flow might deviate from the theoretical ratio. Therefore a correction factor CF was introduced in our model, which can take values between -1 and 1. For $CF = 0$, L/V -ratios equaled theoretical ratios according to Equation (3.4). For positive CF values linearly more lateral flow, for negative CF values linearly less lateral flow was generated in each grid cell; $CF = 1$ resulted in 100%, $CF = -1$ in 0% lateral flow. Analogous to the runoff coefficient model structure, water input on NDP areas contributed to the baseflow generating groundwater linear reservoir (Figure 3.1b).

3.2.1.7 Channel Arrival Time

Two parameters define the overland flow velocity (FV_OF) and shallow subsurface flow velocity (FV_SF). For each grid cell the overland flow path distance was computed. The combination of flow path distance with the appropriate flow velocity resulted in channel arrival times of generated runoff. Because water travel times were much longer on the hillslope compared to the channel, the latter were neglected. This concept is similar to a geomorphologic unit hydrograph concept (Gupta et al., 1980) but assuming different velocities according to the flow pathway (overland or subsurface).

3.2.1.8 Connectivity Modules

Three levels of hillslope - channel network connectivity were tested: no connectivity (Figure 3.1c, left), time-independent connectivity (Figure 3.1c, center), and time-dependent connectivity (Figure 3.1c, right). Without the connectivity module, all delineated runoff generation areas were assumed to be connected to the channel network. The time-independent connectivity module only considered DRP areas which are directly connected to the stream or connected via other stream-connected DRP areas. Connectivity was defined by cell connections along the D8 overland flow paths (O'Callaghan and Mark, 1984). In the time-dependent connectivity module all potential connections were assessed according to the actual generation of runoff at each grid cell in time (Figure 3.1c, right). At each time step, the D8 flow paths were examined from the channel upslope (similar to Lane et al., 2004) with respect to connectivity of grid cells generating runoff at this time step (analogous to Reaney et al., 2007). The proposed connectivity approach was primarily based on the 'connect and react' concept (Bachmair and Weiler, 2010). The combination of runoff generation and connectivity results in areas that generate runoff but are not necessarily connected. In the time-dependent and time-independent connectivity approach it was assumed that water flowing from unconnected grid cells into down slope cells contributed to additional input.

3.2.1.9 Final Model Structures

The combination of the two runoff generation model types (runoff coefficient models and process models; sections 3.2.1.5 and 3.2.1.6) with the three connectivity modules (no connectivity, time-independent connectivity, time-dependent connectivity; section 3.2.1.8) resulted in six different model structures (Table 3.2). All six model structures were used to test the influence of connectivity in DRP models.

Table 3.2: Model structures. The six model structures include two runoff generation model types (runoff coefficients and process models) and connectivity approaches (no connectivity, time-independent and time-dependent).

models	runoff generation modules		connectivity modules		
	runoff coefficient models	process models	no connectivity	time-independent	time-dependent
M1	X		X		
M2		X	X		
M3	X			X	
M4		X		X	
M5	X				X
M6		X			X

3.2.2 Study Site

The model framework was applied to hydrological data from five catchments in the Kootenay Mountains in the East of British Columbia, Canada (Figure 3.2). The catchment areas ranged from 1.03 to 1.64 km² with elevations between 1100 and 2100 m. The slopes and flow path lengths varied substantially among the five catchments (Table 3.3).

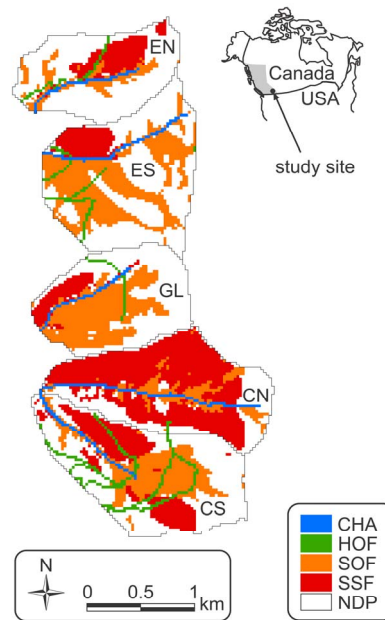


Figure 3.2: Study site. Five catchments: EN = Elk North, ES = Elk South, GL = Glory, CN = Cotton North, CS = Cotton South. The colored areas show a preliminary map of dominant runoff generation processes (CHA = channel interception, HOF = Hortonian overland flow, SOF = saturation overland flow, SSF = shallow subsurface flow, NDP = no dominant runoff generation processes).

Table 3.3: Median gradients and flow path lengths for the five catchments.

	Elk North (EN)	Elk South (ES)	Glory (GL)	Cotton North (CN)	Cotton South (CS)
median gradient to stream [-]	0.23	0.20	0.22	0.31	0.27
median flow path length [m]	337	482	562	314	505

The natural vegetation of the study area was forest, which was dominated by Lodgepole pine (*pinus contorta*). In 2009, 66% of the catchment area was forested and 34% was recently clear cut. In the period from 2005 until 2010, annual precipitation measured at two climate stations in the study area averaged 650 mm. Monthly precipitation maxima of 51 mm (June) and 38 mm (December) and average temperatures of -5.8°C (January) and 17.6°C (August) were observed. Precipitation accumulated as snow in winter; snowmelt in spring dominated the annual runoff regime. During summer, smaller convective rainfall events produce substantial stream flow peaks.

The soil of the study area consisted of a thick compact basal till layer. In some mid slope locations, ablation till mixed with glacio-fluvial sediments could be found. Bedrock outcrops were most often located on ridge tops, along the watershed boundaries, but also along the main divides within the watershed. Colluvial zones were frequently located down slope of bedrock outcrops. Except for the deeply incised stream channels and bedrock outcrops, surficial materials are overlaid by organic rich, 50-100 cm deep soil. The different geomorphological characteristics of the five catchments (e.g. Table 3.3) resulted in substantially different DRP distributions. In Figure 3.2 the DRP areas according to the DRP delineation criteria by Carver et al. (2009) are displayed for thresholds of TOPO = 10, VERT = 3 m, and GTS = 25%.

3.2.3 Data

Two climate stations at different elevations (1390 m, 1780 m) measured precipitation, air temperature, humidity, wind speed and direction, and short wave radiation from 2005 until 2008 with at least hourly resolution. The climate data were spatially interpolated with inverse distance weighting. Snow depth and density were measured with a stratified nested sampling design from 2005 until 2008 (Jost et al., 2009). The snow surveys sampled 19 strata covering different elevations, forest cover, and aspects. In each stratum, 60 snow depth measurements and 12 snow density measurements were taken. In the years 2007 and 2008, the sampling was slightly reduced

(Jost et al., 2009). More detailed descriptions on snow sampling and meteorological data in the region of the five catchments is provided by Jost et al. (2007) and Jost et al. (2009).

The field data were not directly used as input data for the models. However, the data set was used as input data for DHSVM (Distributed Hydrology-Soil-Vegetation model; Wigmosta et al., 1994; Wigmosta et al., 2002). With DHSVM the net water input at the soil surface due to snowmelt and precipitation was simulated considering interception losses. This spatio-temporal data set was carefully benchmarked with field measurements by Jost et al. (2009). The DHSVM output was preferred to the raw data in order to minimize spatial inconsistencies of input data, which could have drastic consequences on connectivity analyses. The DHSVM output was interpolated from a 50 m to a 25 m grid and aggregated from hourly to four-hour time steps. Runoff was measured at hourly time steps at the outlets of the five catchments. Stream flow was continuously simulated for the year 2006 from March 1 until October 31 at four-hour time steps. In 2006, snow melt started in the first week of April with a snow melt maximum in the middle of May. Therefore, it was assumed that on March 1 most SOF and SSF storages were nearly empty.

3.2.4 Storm Types

Temperature and precipitation data from the two climate stations, and the snow water equivalent state variables from the DHSVM simulations, were used to classify the simulation period into four storm types: rain on snow (ROS), snow melt (SME), rain storms (RAI), and unforced periods (UNF, Table 3.4).

Table 3.4: Classification of event types.

Storm Type	Abbr.	Classification
rain on snow	ROS	Precipitation (measured by at least one station) in combination with an average temperature greater than 0°C; snow pack at any location in the five catchments.
snow melt	SME	No liquid precipitation in both stations and an average temperature greater 0°C, but input simulated by DHSVM.
rain storms	RAI	Precipitation (measured by at least one station); average temperature greater than 0°C; no snow pack at any location.
unforced periods	UNF	All time periods other than ROS, SME, or RAI.

One storm type was assigned to each four hour time step of the analyzed period. However, a certain influence from one storm event on the runoff of the following time periods was assumed. This influence period was assumed to last two days for ROS, SME and RAI. These durations of the single storm types were combined under the assumption that ROS dominated SME, RAI and

UNF; UNF was subordinate to all other storm types, and the combination of SME and RAI resulted in ROS. This resulted in a final temporal distribution of storm types from March until October 2006 of 22.1% ROS, 14.5% SME, 30.8% RAI, and 32.6% UNF.

3.2.5 Model Benchmarking

To assess whether the incorporation of connectivity modules improves DRP models we looked at three aspects of potential model improvement -- model fit, prediction uncertainty, and parameter feasibility -- which are explained in the next sections.

3.2.5.1 Objective Functions for the Model Fit Evaluation

The Nash-Sutcliffe Efficiency (NSE, Nash and Sutcliffe, 1970) was used to evaluate the general runoff dynamics focusing on peak flow. The evaluation of the runoff volume simulations was based on a mean absolute error. To facilitate comparison among the catchments, the mean absolute error was standardized by the maximum observed runoff for each catchment (MAM, Equation 5):

$$MAM = \frac{\frac{1}{n} \sum_{i=1}^n |Q_{obs,t} - Q_{sim,t}|}{Q_{obs,max}} \quad (Eq. 3.5)$$

3.2.5.2 Prediction Uncertainty

Model prediction uncertainty was derived with an extended GLUE (generalized likelihood uncertainty estimation) approach (Beven and Binley, 1992). For each model structure and catchment, 10,000 simulations were executed based on uniform distributions within *a priori* defined sampling ranges (Tables 3.5 and 3.6). In our study the uncertainty ranges represent the runoff ranges of the best 10% of the runoff simulations separately for NSE and MAM. Prediction uncertainty was quantified with a relative range at peak, which was defined as the uncertainty range at peak, divided by the observed runoff at peak, to allow for comparisons between watersheds.

3.2.5.3 Parameter Feasibility

Another criterion to evaluate model improvement was whether optimized parameters fall within previously defined feasible parameter ranges. Fuzzy parameter ranges were selected; outside the *a priori* defined outer limits (lower feasibility limit 2 and upper feasibility limit 2 in Table 3.5 and 3.7) the feasibility was considered to be poor (feasibility value = 0). Inside inner limits the parameter feasibility equaled; between lower and upper limits the feasibility was linearly interpolated.

Table 3.5: Sampling and feasibility ranges of runoff coefficient models (M1, M3, M5).

Parameter Group	Parameter Names	Units	Lower Sampling Limit	Lower Feasibility Limit 2	Lower Feasibility Limit 1	Upper Feasibility Limit 1	Upper Feasibility Limit 2	Upper Sampling Limit
areas	TOPO	[-]	8.0	9.0	9.5	10.5	11.0	13.0
	VERT	[m]	0	1	2	5	10	20
	GTS	[%]	15	20	23	26	32	40
runoff generation	CHA_RC	[-]	0.01	0.25	0.35	0.65	0.70	0.99
	HOF_RC	[-]	0.01	0.20	0.30	0.70	0.80	0.99
	SOF_RC	[-]	0.01	0.20	0.30	0.70	0.80	0.99
	SSF_RC	[-]	0.01	0.25	0.35	0.75	0.85	0.99
	K_DP	[1/h]	0.0001	0.0005	0.001	0.004	0.008	0.05
routing	OF_FV	[m/h]	5	10	15	40	50	80
	SF_FV	[m/h]	1	4	6	8	10	30

Table 3.6: Sampling and feasibility ranges of process models (M2, M4, M6).

Parameter Group	Parameter Names	Units	Lower Sampling Limit	Lower Feasibility Limit 2	Lower Feasibility Limit 1	Upper Feasibility Limit 1	Upper Feasibility Limit 2	Upper Sampling Limit
areas	TOPO	[-]	8.0	9.0	9.5	10.5	11.0	13.0
	VERT	[m]	0	1	2	5	10	20
	GTS	[%]	10	20	23	26	32	40
runoff generation	CHA_TH	[-]	0.01	0.25	0.35	0.65	0.70	0.99
	HOF_TH	[mm/h]	0.01	0.05	0.1	0.5	1.0	3.0
	SOF_UPLIM	[mm]	10	20	40	100	150	300
	SSF_UPLIM	[mm]	10	20	40	100	150	300
	SSF_CF	[mm]	-0.99	-0.70	-0.30	0.30	0.70	0.99
	K_DRP	[1/h]	0.001	0.005	0.01	0.04	0.08	0.2
	K_DP	[1/h]	0.0001	0.0005	0.001	0.004	0.008	0.02
routing	OF_FV	[m/h]	5	10	15	40	50	80
	SF_FV	[m/h]	1	4	6	8	10	30

3.2.6 Computational Methods

3.2.6.1 Optimization with GENOUD

The two objective functions were optimized separately for all combinations of six model structures, five catchments and four storm types. The optimization was done with GENOUD (Genetic Optimization Using Derivatives; Sekhon and Mebane Jr, 1998; Mebane Jr and Sekhon, 2009).

3.2.6.2 Sensitivity Analysis with Sobol and Morris

The sensitivity analysis methods of Sobol (1993) and Morris (1991) were used to assess parameter sensitivity for the different model structures, watersheds, storm types and objective functions. The method of Sobol was applied based on the algorithms by Saltelli (2002). The method of Morris was applied according to the algorithms by Campolongo et al. (2007); the total sensitivity index was estimated with μ^* , which is an estimate of the mean of the distribution of the absolute values of the elementary effects. Detailed methods descriptions and a comparison of μ^* with traditional variance-based measures can be found in the publications by Pujol (2009) and Campolongo et al. (2007).

3.3 Results

3.3.1 Contributing Areas

Considering connectivity in dominant runoff generation processes models had substantial effects on the shapes and sizes of the contributing areas.

3.3.1.1 Shape of Contributing Areas

Considering connectivity resulted in the runoff generation areas being redistributed toward locations along flow paths with continuous connections to the channel network for all catchments and models. The shapes of the connected DRP patches were similar for different catchments and storms, because the rearrangement followed the model assumption of connectivity along flow paths. Figure 3.3 shows an example for the CS catchment using model M4. Patches without continuous connection to the channel, which were mainly HOF and SSF areas, were not connected.

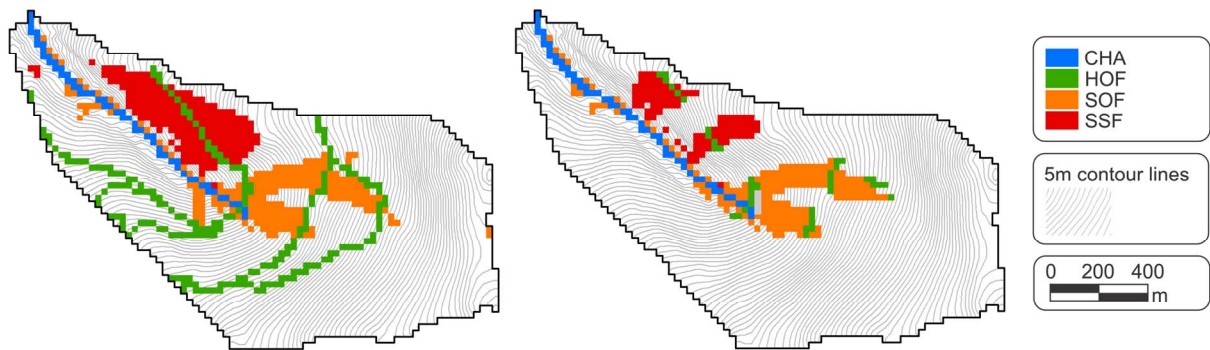


Figure 3.3: Effect of connectivity on the shape of contributing areas. Contributing areas without (left) and with connectivity (right). DRP areas (CHA, HOF, SOF, and SSF) of the CS catchment are displayed according to model M4 with optimized parameters for on rain on snow events based on the Nash-Sutcliffe efficiency.

3.3.1.2 Size of contributing areas

The sizes of DRP areas varied considerably across catchments and storms. Figure 3.4 presents the effect of connectivity on DRP areas. For each watershed the contributing area according to different dominant processes is shown for runoff coefficient (left column) and process models (right column). The rows of the graphs represent the four storm types and two objective functions.

The biggest differences in the size of DRP area were found among different catchments and model types. For models with time-dependent connectivity modules (models 5 and 6), temporal averages of contributing areas are displayed in Figure 3.4. This size of contributing areas was in agreement with the model set up, in which shallow subsurface flow was generated in steep catchments (e.g. CN catchment), and more saturation overland flow occurred in flat catchments (e.g. ES and GL).

In general, the DRP areas of the process models were slightly larger than for the runoff coefficient models. Moreover, the variability of the size of contributing areas among different storm types is higher for the runoff coefficient models than for the process models. For example in the CS catchment the contributing areas of the runoff coefficient models decreased considerably from rain on snow (ROS) to water input periods with less volume and intensity (SME, RAI, UNF). However, the contributing areas of the process models remained almost constant for different water input periods. Considering connectivity resulted in a median reduction of the contributing areas of approximately 10%. No consistent trends of the connectivity-based area reduction were observed for different models, storm types, catchments, or objective functions.

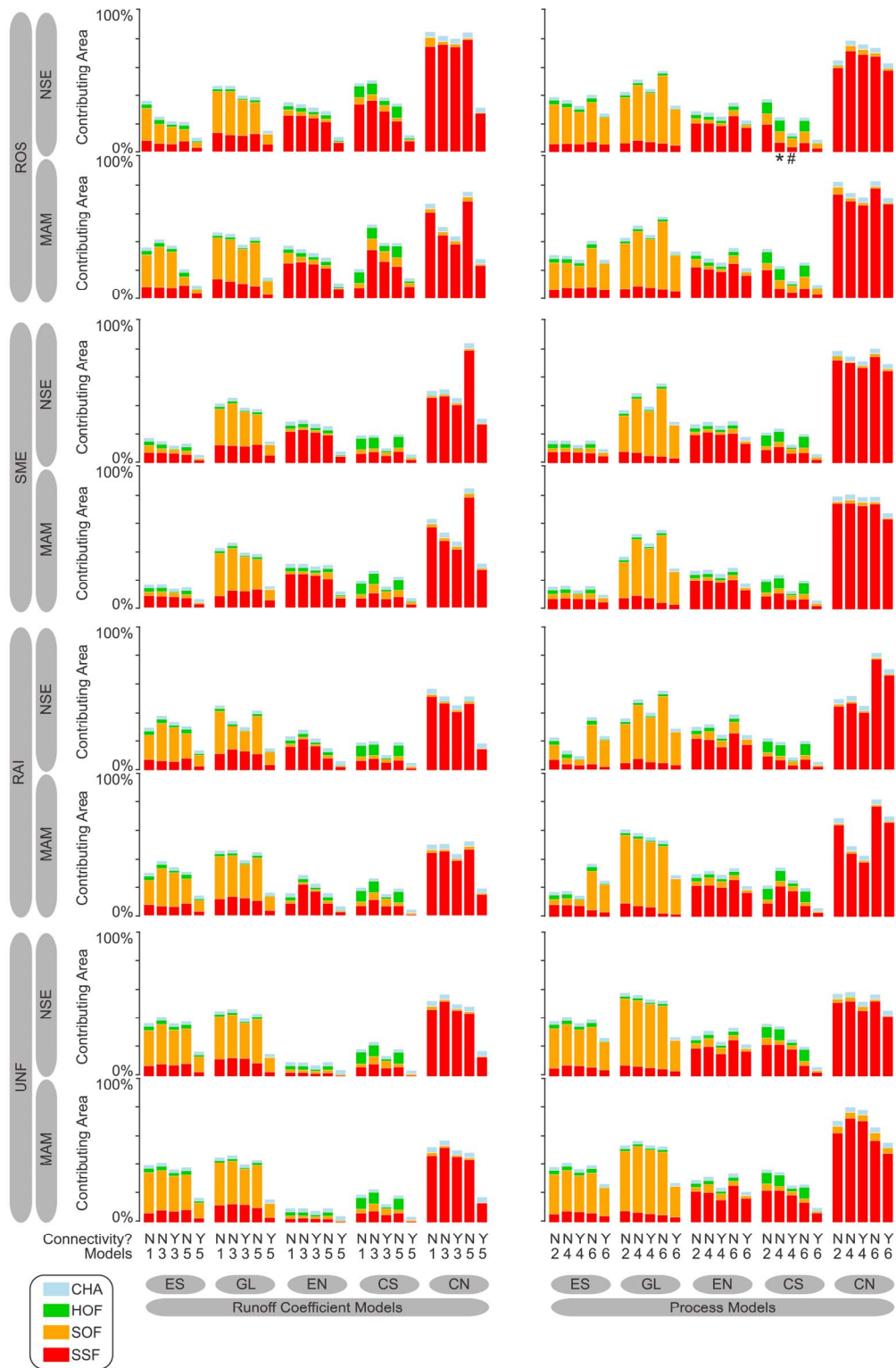


Figure 3.4: Effect of connectivity on the size of contributing areas. Contributing areas per catchment areas are shown for different models, catchments, events, and objective functions. For models 3, 4, 5, and 6 the contributing areas are shown for the initial DRP delineation (connectivity = N) and after the connectivity incorporation (connectivity = Y). The contributing areas in Figure 3.3 correspond to the symbols * (Figure 3.3, left) and # (Figure 3.3, right).

3.3.2 Model Benchmarking

3.3.2.1 Model Fit

In general, the model fits according to NSE and MAM were reasonable. The best model fits for the different objective functions, catchments, models, and storm types are shown in Figure 3.5. The color scale of the larger squares represents the objective function. The smaller squares indicate if considering connectivity improved (white squares) or impaired (black squares) the fit.

Model fits based on NSE varied more among catchments than model type or event type. The best fits were observed for EN and CS, and the worst for CN (Figure 3.5). In addition to the variations among the catchments, the water input type was found to be important for NSE. While only minor NSE differences appeared between rain on snow, snow melt, and rain storms, the NSE was notably worse in periods of unforced runoff generation. NSE values of the process models were slightly better than the NSE of the runoff coefficient models for snow melt, rain storms, and unforced runoff generation; however the process models were considerably better than the runoff coefficient model for rain on snow periods.

The best models according to the MAM objective function also varied among different catchments, but not as clearly as for the NSE. Furthermore, the catchment dependence differed between model types. For the runoff coefficient models the MAM was slightly worse for ES compared to other catchments. However, for process models the MAM improved notably for steeper catchments (Figure 3.5). The largest differences between runoff coefficient and process models were observed for rain on snow events.

The influence of connectivity was assessed for different factors:

- Storm types: For rain events connectivity improved the model fit only in 30% (ratio of white squares for RAI) of the assessed situations. This was clearly less compared to the three other storm types.
- Catchments: The two catchments with the smallest relative contributing areas (EN and CS) exhibited not only the best NSE values but also substantially higher amount of situations with connectivity-based fit improvement compared to the other watersheds. However, for MAM, the differences between catchments were smaller than for NSE.

- Model types: Connectivity more often increased the objective function for the process models compared to the runoff coefficient models. The improvement ratios (i.e. ratios of white squares in Figure 3.5) for the two model types are also displayed in Figure 3.11 and then further discussed.

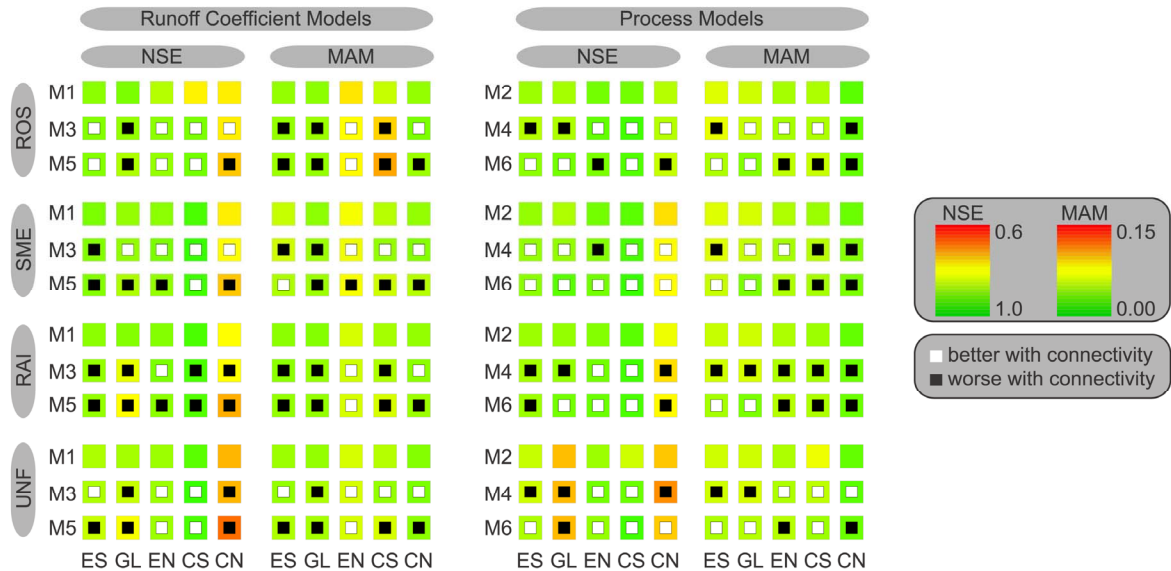


Figure 3.5: Model fit. The model fit according to the Nash-Sutcliffe Efficiency (NSE) and the relative mean absolute error (MAM) is shown for runoff coefficient models (M1, M3, M5), process models (M2, M4, M6), four water input types (ROS, SME, RAI, UNF) and five catchments (ES, GL, EN, CS, CN).

3.3.2.2 Prediction Uncertainty

The prediction uncertainty was estimated with the relative range at peak. This measure exhibited the highest values for rain storms, particularly for the GL, ES, and CN watersheds. The process models had higher prediction uncertainty than the runoff coefficients. However, this cannot be compared directly, since the relative range at peak depended on the sampling ranges, which differed for the two model types. In more than 60% of the assessed situations in Figure 3.6, considering connectivity decreased the prediction uncertainty. However, different effects of incorporating connectivity on prediction uncertainty were only found across catchments (but not objective functions, storm or model types): prediction uncertainty decreased more for steep than for flat catchments if connectivity was considered.

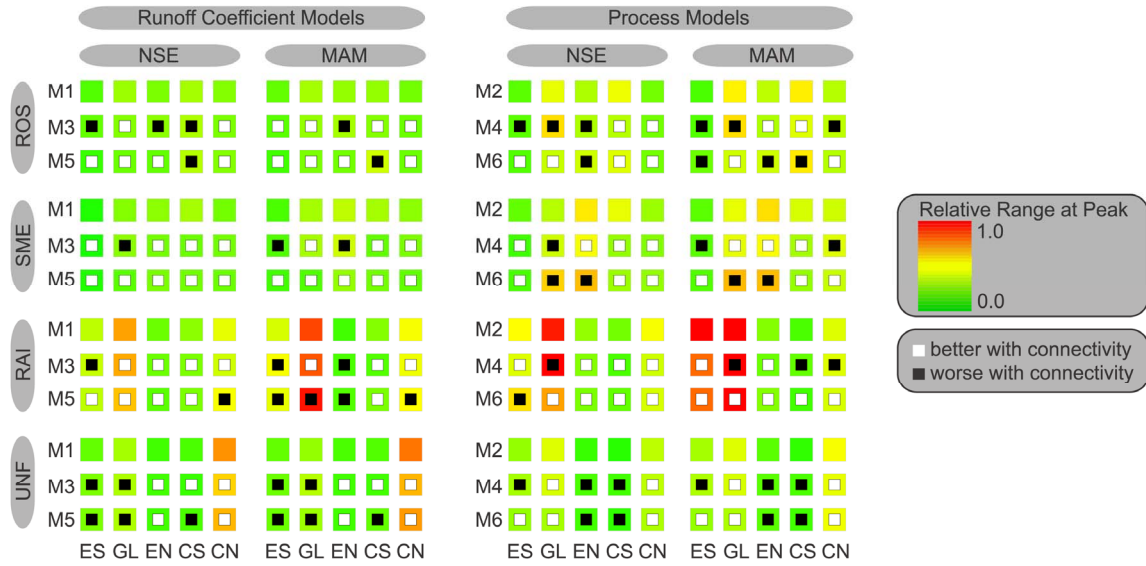


Figure 3.6: Prediction uncertainty. The models (M1 to M6) were optimized with two objective functions (NSE, MAM) for four water input types (ROS, SME, RAI, UNF) and five catchments (ES, GL, EN, CS, CN).

3.3.2.3 Parameter Feasibility

Only sensitive model parameters were considered in the feasibility analysis. Therefore, the next section presents the results of the sensitivity analyses, which were used for the feasibility analysis. Parameters with total sensitivity indices exceeding 50% of an equal sensitivity distribution across parameters (e.g. 0.2 for five parameters) were considered to be sensitive. To allow for comparisons across connectivity approaches and catchments, the same parameter sets were selected for all catchments within runoff coefficient and process models. Parameters were considered sensitive if they were assigned to be sensitive by both methods in at least one catchment. As visualized in Figure 3.7, the three parameters TOPO, GTS, and SF-FV were considered sensitive in runoff coefficient models. The sensitive parameters of process models were TOPO, GTS, SSF-UP, and SSF-CF.

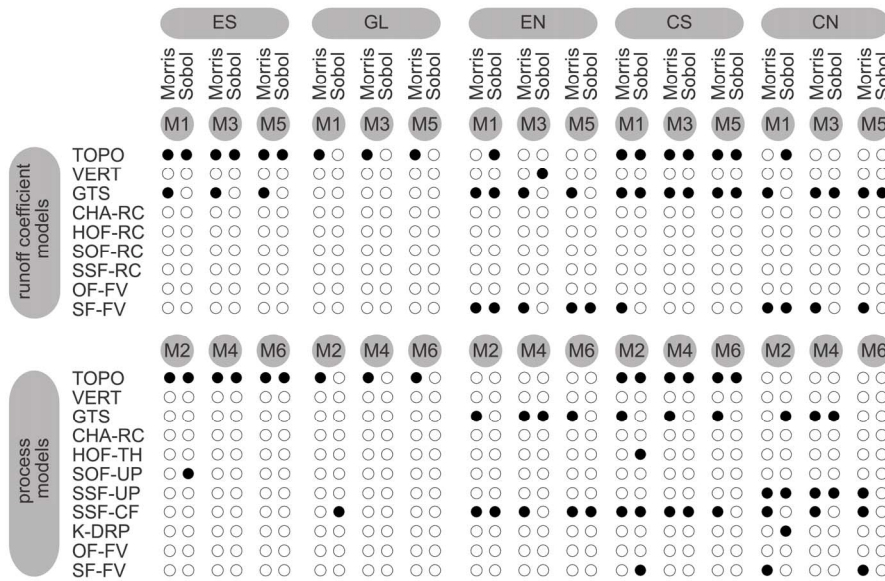


Figure 3.7: Parameter sensitivity. Sensitive parameters of runoff coefficient models (top row) and process models (bottom row) for different catchments (CN, CS, EN, GL, ES) and sensitivity concepts (Morris, Sobol) are displayed with black circles.

The results of the parameter feasibility analysis are presented in Figure 3.8. Poor parameter feasibility was found for runoff coefficient models applied to steep catchments (CS and CN). For the process models, a clear dependency on the catchment could not be seen. The influence of connectivity in Figure 3.8 is shown by the feasibility of optimized parameters of models without (M1, M2) or with connectivity (M3, M4, M5, M6). In the majority of situations connectivity improved the parameter feasibility. The amount of simulations where connectivity had a positive effect on the parameter feasibility was 15% higher than where connectivity had a negative effect. The connectivity-based improvement was clearly higher for process than for runoff coefficient models.



Figure 3.8: Parameter feasibility. The parameter feasibility of the optimized parameters was rated based on previously defined fuzzy parameter ranges.

3.3.3 Differences Between Runoff Generation Processes

Runoff generation areas of the dominant processes HOF, SOF, and SSF were simulated to be located in areas with process-specific characteristics. Our results demonstrated that the effects of considering connectivity interfered with the process-specific locations. Differences between runoff generation processes can be seen in Figure 3.9, which shows the empirical cumulative distributions F_{emp} of the durations of connectivity for the grid cells of each process. The durations were standardized by the duration of the simulation (abscissas in Figure 3.9). The empirical distributions were calculated with Weibull plotting positions. For the model structures

with time-independent connectivity modules (models 3 and 4) only the proportions of connected cells can be identified in Figure 3.9, but no temporal variability. For time-dependent connectivity approaches (models 5 and 6) both spatial and temporal variability is displayed.

Substantially fewer proportions of HOF cells were connected to the channel network than SOF or SSF cells, irrespective of model or catchment. However, the differences between the DRPs were marginal for model 5. The distributions varied more gradually for model 6; abrupt distributions appeared only for small amounts of grid cells per process (e.g. the CN catchment only contained six HOF grid cells).

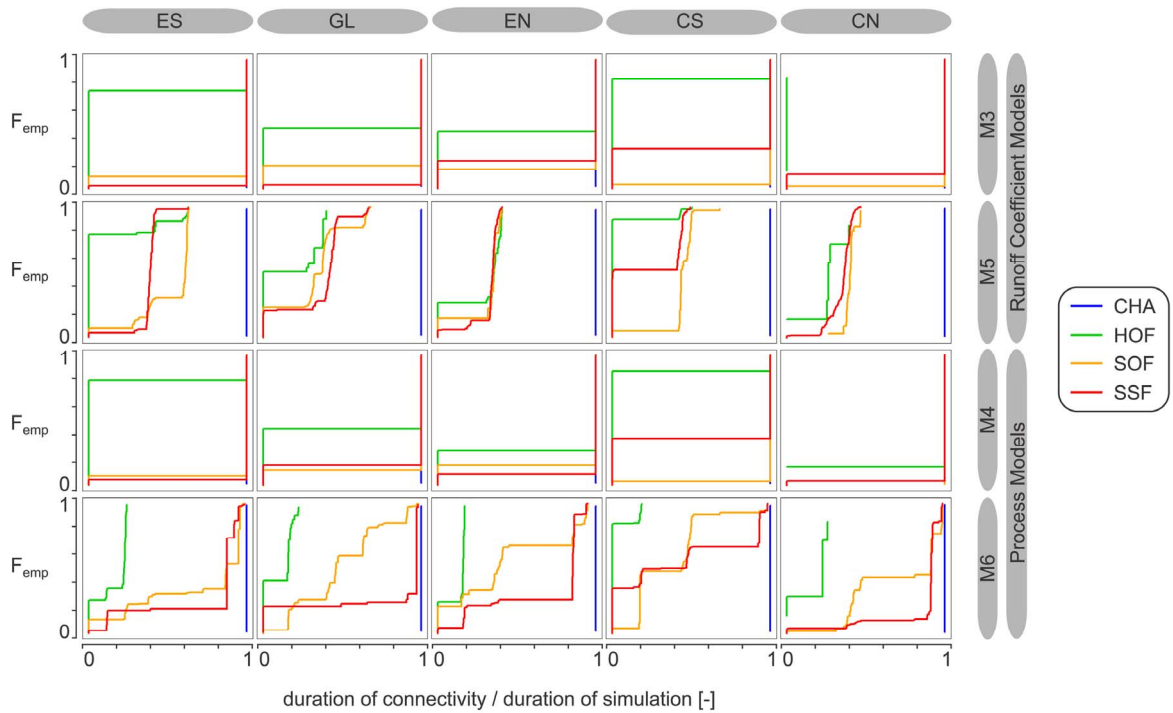


Figure 3.9: Temporal connectivity distribution. Empirical cumulative distributions F_{emp} of the durations of connectivity for the grid cells of each process (CHA, HOF, SOF, SSF) are shown for different catchments (ES, GL, EN, CS, CN) and models. The parameters were optimized based on the Nash-Sutcliffe Efficiency over the entire simulation period.

3.3.4 Computational Costs of Connectivity

The computation time per model run is of importance for a useful simulation tool. In general, the process models were slower than the runoff coefficient models (Figure 3.10). However, the time-independent connectivity module did not cause any increase of the simulation duration. The time

to assess connectivity and the reduced amount of processed cells cancelled out. Models with time-dependent connectivity modules were slower. The increase of the simulation time was based on the length of the simulation period. Furthermore, the simulation time highly depended on the amount of DRP cells within each catchment.

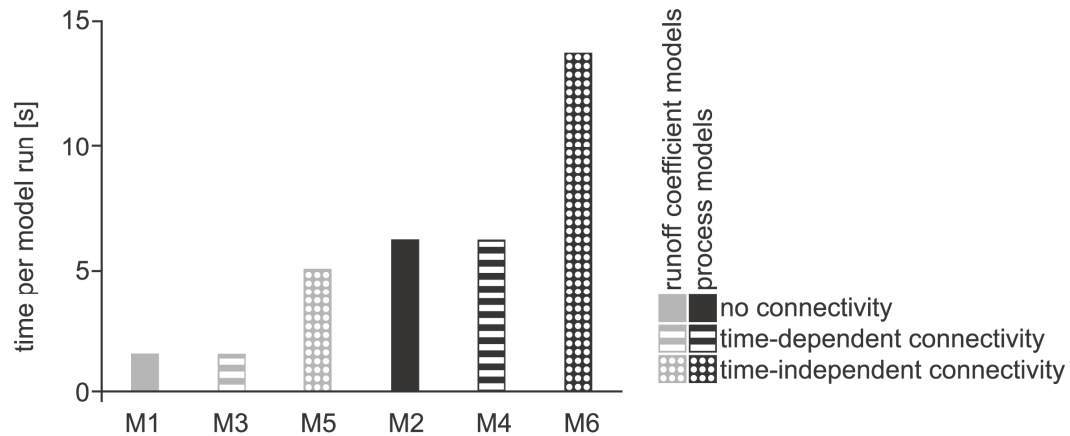


Figure 3.10: Simulation times of different models for the GL catchment.

3.4 Discussion

3.4.1 Contributing Areas

3.4.1.1 Shape and Size of Contributing Areas

In order to simulate similar runoff volumes, the size of the contributing areas should theoretically remain constant with or without connectivity. For constant runoff volume simulations the model parameters should be optimized in such a manner that DRP areas were only redistributed along flow paths at similar total size. However, the runoff generation areas were often reduced when connectivity was considered (Figure 3.4). This reduction partly originated from the rather narrow ranges of area parameters TOPO, VERT, and GTS (Table 3.6). The limits of TOPO and VERT were selected based on Rosin et al. (2010), who compared terrain analysis models with field data in four regions in British Columbia. The GTS limits were estimated based on experience from experimental studies of shallow subsurface flow generation (McGuire et al., 2005). The rather narrow area parameter ranges were selected to avoid parameter interaction and overcompensation effects with the runoff generation parameters.

3.4.1.2 Contributing Areas for Different Model Types

The contributing areas according to runoff coefficient and process models could reveal the differences in flexibility between the two model types. The DRP areas in process models remained almost constant for different storm types, because a more realistic runoff generation was achieved with flexible process models for different storm types; the reservoir approach was in particular helpful to incorporate antecedent conditions. In some catchments, the runoff coefficient models could only achieve realistic runoff reactions by drastic variations of the runoff generation areas. Moderate variations of contributing areas have been observed in British Columbia (Rosin et al., 2010). Variations in areas of more than 200% (e.g. for the CS catchment; the ratio of ROS-area compared to SME-areas for runoff coefficient models) were unrealistic and considered inconsistent with the variable contributing area concept (Hewlett and Hibbert, 1967).

3.4.2 Model Benchmarking

3.4.2.1 Catchments

The assessment of model fit and uncertainty with the Nash-Sutcliffe efficiency showed that runoff was reasonably well simulated with DRP models for the two catchments EN and CS. High NSE values and uncertainty ranges corresponded to small contributing areas (Figures 3.4 and 3.5). The smaller DRP areas of EN and CS reflected the lower specific discharge at peak flow of 0.34 mm/h (EN) and 0.22 mm/h compared to the other catchments (CN; 1.35 mm/h, ES 0.54 mm/h, GL 0.55 mm/h). Peak flows in CN, ES, and GL were mainly simulated based on one dominant process (SSF for CN, SOF for ES and GL).

However, we hypothesize that not the size of the contributing area (Figure 3.4) but rather the distribution among DRP processes might have caused the differences in model fit. DRP models simulate predominantly SSF in steep and SOF in flat watersheds. The observed hydrograph followed this hypothesis at peak flow, but not for several sub-peaks. Consequently, the DRP models were more successful in watersheds with fairly equal distributions of HOF, SOF and SFF, as in the CS and EN catchments (Figure 3.4). On the other hand, the simulations were worse for the ES and GL catchments, which were dominated by SOF and CS by SSF. The predictions in these watersheds were highly sensitive to parameters of the dominating processes (Figure 3.7). The NSE-optimized simulation of these catchments was predominantly orientated on the main peak, but not on sub-peaks. Consequently, the best DRP simulations might be

achieved for watersheds with fairly homogeneous process distributions. We conjectured that the size of contributing areas and the runoff level played a minor role (e.g. higher likelihood of discharge measurement errors at peak flow for watersheds with high runoff).

3.4.2.2 Objective Functions

The better MAM values for steeper catchments could, to some extent, be explained by the fact that runoff generation in steep watersheds agrees better with mechanisms inferred from the D8 flow path algorithm than in flat areas, where the flow gradients are more ambiguous. This resulted in realistic simulations of generated runoff volumes and fairly good MAM values over the entire simulation period. Irrespective of model and objective function, the lowest NSE values were achieved for periods of unforced runoff generation, which lacked hydrograph peaks. To some extent the low NSE values can be explained by the low flow variance and subsequent high relative error variance during periods of unforced runoff generation.

3.4.2.3 Model Types

The biggest differences in model fit were observed for rain and snow periods. For both objective functions considerably better simulations were achieved with process models. This was a strong argument to apply process models, since the motivation of our study was to assess peak flows, which are often generated by rain on snow events in British Columbia.

3.4.2.4 Connectivity and Model Type

Effects of connectivity on model fit, prediction uncertainty, and parameter feasibility were only found among a few factors. The strongest effect was observed between model types (Figures 3.5, 3.6, and 3.8). Assessment of model fit and feasibility showed a higher benefit of connectivity in process models than in runoff coefficient models (Figure 3.11). The prediction uncertainty was not directly comparable between model types due to its dependence on parameters which differed substantially between model types. The parameter feasibility of runoff coefficient models was low in steep catchments due to parameter compensation effects. The values of the GTS threshold parameter were too high in the steep watersheds. This resulted in an underestimation of the size of contributing areas; due to the limited properties of runoff coefficients, reasonable simulations were only possible with underestimated contributing areas. Consequently, we hypothesized that it made more sense to incorporate connectivity modules in the more realistic process models. The process models exhibited less parameter compensation

effects (e.g. an increase of area parameters due to a lack of flexibility in runoff generation parameters). The connectivity modules provided additional support to the process models especially with respect to considering antecedent conditions.

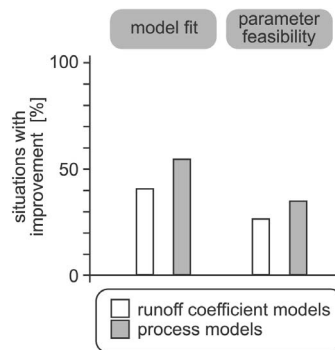


Figure 3.11: Improvement due to connectivity. The percentage of improvement of the model fit and the parameter feasibility is summarized across runoff coefficient and process models.

3.4.2.5 Storm and Connectivity Types

We hypothesized that the antecedent conditions before events in British Columbia might cause simulation differences between rain on snow and summer rain storm. Peak flow is often generated in late snow melt periods, in which snow melt generates high variability of dry and wet patches. In these periods simulations improved when connectivity was considered (Figure 3.5). However, most locations of the study area were dry. Field observations showed that wet areas were mainly found close to streams. Consequently, connectivity modules did often not improve the simulations for summer rain storms (Figure 3.5). The simulations were not substantially better with time-dependent compared to time-independent connectivity modules (Figures 3.5, 3.6, and 3.8). Due to the lower computational costs time-independent connectivity modules may be preferred.

3.4.3 Differences between Runoff Generation Processes

Considerable differences of connectivity over time appeared between runoff generation processes (Figure 3.9), which can mainly be explained by the spatial DRP delineation criteria. The HOF generating roads were predominantly located perpendicular to flow paths (Figure 3.3). However, the patches of SOF and SSF were often found along flow paths. In runoff coefficient models, simulation differences among DRPs occurred only due to different DRP area locations and slightly varying runoff coefficient values. Consequently, the temporal

connectivity distributions of model 5 also only exhibited small differences across processes (Figure 3.9). In contrast, the process models showed bigger differences between processes. The distributions of model 6 reflected more realistic characteristics of the different runoff generation processes (Figure 3.9):

- Since most HOF areas were located upslope and perpendicular to the flow paths, no HOF cell was connected more than 30% of the time in any catchment. The HOF cells reacted similar, which was expressed by steep, abrupt distributions in Figure 3.9.
- In reality SOF areas are often located next to streams or along flow paths. During storm events saturated areas expand and shrink according to the corresponding soil saturation deficit. Therefore, gradual and not abrupt temporal distributions of connected SOF areas are expected theoretically. In Figure 3.9, gradual distributions can be found for SOF, particularly for the SOF-dominated GL catchment. These SOF distributions are consistent with realistic process representation.
- SSF areas exhibited patchy shapes and were often located along flow paths, and were often found upslope. These two facts resulted in step-like connectivity distribution functions. For example in the ES catchment, approximately 70% of the SSF areas were connected for more than 80% of the time; however, 10% of the SSF cells were not connected more than 20% of the time.

Across all dominant runoff generation processes, temporal connectivity indicated more realistic process simulations by process models compared to runoff coefficient models

3.4.4 Evaluation of the Model Structures

3.4.4.1 Area Delineation

The introduced model structure distinguished between areas which did or did not contribute to runoff generation. In the process model structures the contributing areas could be reduced to flexibly cope with the actual runoff generation. However, it should be noted that the area adjustment was only related to initially delineated contributing areas. Consequently, it was not possible that classified areas with no dominant runoff generation process (NDP) could contribute to runoff generation after long lasting water input. This fact is apparent in Figure 3.3, in which SSF cells were not connected to the stream network due to single NDP cells. It has to be clarified that all area delineation criteria of DRP models in the past (e.g. Scherrer and Naef, 2003) only reflect runoff generation only for water input (precipitation or snowmelt) intensity and duration.

It can be presumed that all NDP cells would contribute to runoff, for example, for a storm with higher return period. This entails that delineation of runoff generation areas implicitly based on ranges of water input characteristics (Faeh et al., 1997).

3.4.4.2 Process Models

The simulated runoff generation in DRP models is based on a single process; however in reality multiple processes contribute to runoff generation. Consequently, parameters of one process compensate for the lack of other processes. For example, the size of the linear SOF-reservoirs was possibly overestimated to compensate for omitted infiltration constraints of the soil surface. The findings of this study are restricted to runoff generation due to topographical features. Ignoring soil structure and texture, as well as preferential flow paths, probably leads to overly homogeneous subsurface flow responses.

3.4.4.3 Connectivity

The connectivity modules of this study followed the approach by Lane et al. (2004), which is based on the concept of connectivity along theoretical, topography-driven flow paths. Due to the usually shallow soil depths to bedrock in British Columbia, the assumption of D8 flow path behavior probably does not deviate considerably from reality. Furthermore, we assumed that connectivity can occur between different processes (e.g. SSF areas can be connected to the channel with SOF areas); this comprehensive connectivity approach across processes is also explained by the shallow soil depths in British Columbia. It is in agreement with Anderson et al. (2008) who observed that flow in high-permeability features interconnected via saturated soil patches. Moreover, the lack of soil data did not allow any well-founded case differentiation of connectivity between processes. We would like to point out that due to lack of data we did not consider small scale characteristics, which could have a substantial influence on connectivity in reality (Loos and Elsenbeer, 2009). Hillslope – channel network connections are often driven by threshold dynamics (e.g. Tromp-van Meerveld and Weiler, 2008; Jencso et al., 2009) which can be simulated with our connectivity approach. In particular the model structure M6 managed to reproduce non-stationary connectivity dynamics similar to those outlined by Bracken and Croke (2007).

3.4.5 Future Research

The connectivity concept represents a promising tool for precipitation runoff simulations. However, model simulations should be evaluated with field data. Such evaluations should not be purely based on runoff, but should also include spatial provenance indicators such as stable isotopes, geochemistry or pollutants. Furthermore, internal observations such as soil water content, water table response of the saturated zone, and groundwater should be used for the model evaluation as well. We believe that explicit incorporation of the volumes to breakthrough concept (Bracken and Croke, 2007) in connectivity simulation models could provide a tool to improve both system understanding and connectivity predictions based on antecedent conditions. However, we suggest using the volume to breakthrough not as a stand alone measure for connectivity quantification but in combination with spatial distributions of connectivity durations.

3.5 Conclusions

This study introduced a framework to incorporate and evaluate connectivity modules in spatially distributed, process-based hydrological models. The framework consisted of two model types differing in complexity. Within these two models, different connectivity approaches were assessed for five catchments, different storm types, and objective functions. The results provided evidence that incorporating connectivity in a dominant runoff generation process model leads to a spatial redistribution of contributing areas toward locations along flow paths. The findings suggest that the model quality can be improved with connectivity modules; however, it makes more sense to consider connectivity in fairly realistic process models than in simple runoff coefficient models. This study supports the hypothesis that Hortonian overland flow is more affected by connectivity than other runoff generation processes. The findings suggest that considering time-independent connectivity exhibit less realistic temporal distributions of connectivity than time-dependent approaches; however, the former are available at low computational cost. Consequently, we suggest using time-independent connectivity modules for simulations with low temporal resolution (e.g. daily time steps), for which exact simulations of expansion and shrinking are of minor relevance. For high temporal resolution simulations (e.g. hourly data and small watersheds), time-dependent approaches may be preferred.

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4 PREDICTION OF CATCHMENT RESPONSE WITH DOMINANT RUNOFF GENERATION PROCESSES¹

4.1 Introduction

The way how a catchment responds to precipitation or snowmelt has major impacts on aquatic ecosystems (Townsend, 1989). The dynamics of streamflow, which can be seen as catchment response to water input, have been accepted as a first order control on the organization and function of aquatic ecosystems (Monk et al., 2007). Flow dynamics and change rates influence abiotic habitat characteristics and affect ecosystem relationships (Richter et al., 1997). Furthermore, the catchment response is a key factor in flood management (Carpenter et al., 1999).

4.1.1 Factors of Catchment Response to Water Input

Certain factors of a catchment might control or dominate the dynamics of the streamflow at the outlet; they are called catchment response factors (Shah et al, 1996). Finding controlling factors to predict catchment response has been a key objective in hydrology research for many decades. Hewlett and Hibbert (1967) found soil depth and permeability, average slope, slope length, storm size and frequency, mean annual precipitation, and landuse to be major factors to predict response of humid watersheds. Catchment characteristics with relevance for the dynamic of runoff generation have been assigned to catchment functions (Black, 1997). Wagener et al. (2007) differentiated catchment functions into partition, storage and release. This differentiation could help to understand and consequently predict catchment response behavior.

Often specific aspects of catchment response factors have been predicted: Bell and Moore (2000) investigated watershed partitioning for different precipitation types with distributed models. The effects of landuse changes on catchment response have been simulated by Storck et al. (1998). Many studies have elucidated the influence of watershed response on water quality (e.g. Boyer et al., 1997). Furthermore, water transit times differences among catchments have been studied (Hrachowitz et al., 2009). The knowledge of factors dominating catchment response has been

¹ A version of this chapter will be submitted for publication. Rosin, K., Weiler, M., 2010. Prediction of catchment response with dominant runoff generation processes.

used for scenario analyses. For example, Drogue et al. (2004) simulated variations of watershed responses due to climate change scenarios.

4.1.2 From Catchment Response Prediction to Catchment Classification in Ungauged Basins

Comprehensive approaches to predict catchment response have emerged from ecological studies. The need of comprehensive flow regime predictions for conservation decision has been demonstrated by Richter et al. (1996). Poff et al. (1997) called for wide-ranging methods of flow regime assessment. Snelder et al. (2005) predicted a variety of ecologically relevant stream flow characteristics with catchment properties. They formulated their findings into a river environment classification system. Wagener et al. (2007) evaluated and integrated different hydrological classification approaches, and pointed out the need for predictive power of classification. Classification should not solely be used to categorize watersheds based on similarity, but with respect to flow and process predictions (Snelder et al., 2009). Sivakumar (2004) suggested classifying watersheds into low- and high-order dominant process systems; later, a dominant process based catchment classification framework was introduced (Sivakumar, 2008). Furthermore, classification should allow catchment intercomparisons across different climates and spatial scales (McDonnell and Woods, 2004).

Catchment classifications with predictive power have a strong potential to be applied in ungauged basins (Sivapalan et al., 2003). Despite the fact that stream flow is a watershed integrating measure (Lake, 2007), a powerful prediction framework for ungauged basins should preferably be founded on spatially distributed process understanding (Mc Donnell et al., 2007). Therefore, conceptual models can improve system understanding and uncertainty reduction of predictions in ungauged basins (e.g. Yadav et al., 2007).

4.1.3 Research Gaps and Study Outline

Catchment classifications often rely on detailed watershed information such as morphological characteristics (e.g. Lee et al., 2006). However, process approaches are often neglected in ungauged basins. Fenicia et al. (2008) concluded that models were often oversimplified in situations with scarce data. Specifically, types of data (such as topography, water bodies, landuse climate) vary considerably depending on the location of the watersheds, region or country: Topographic data are available worldwide (Rabus et al., 2003). Stream network data, if not

available from mapped data sources, could be estimated from topographic data, although the extent and density of the simulated stream network can be fairly uncertain (McMaster, 2002). However, landuse and accurate spatial distribute climate data are often not available. Despite different levels of data availability, only a few studies have investigated the predictive performance of hydrological models under data shortage (e.g. Grayson et al., 2002). Specifically, it has not been assessed if the delineation of runoff generation processes can improve the catchment response predictions in ungauged basins. Furthermore, it has not been studied if process-based runoff predictions have value in absence of data about central catchment response factors such as soil or climate.

In this study, we then investigated the power of the spatial distribution of dominant runoff generation processes (DRP; see Carver et al., 2009) for catchment response predictions under different levels of data availability. This study investigated how much of the variability in runoff response among catchments can be explained by watershed characteristics using multiple linear regression analysis. We incorporated contributing area concepts in stream flow predictions, which could be of substantial interest for ecological studies, since the catchment response forecasts can be underlain with the source areas of runoff. This is of particular relevance for studies dealing with the effects of landuse change und water quantity and quality. Our concept may enhance a comprehensive hydrological-ecological framework for catchment response effects and environmental needs (Poff et al., 2009), in particular for data-scarce basins. In contrast to Sivakumar (2008) the identification of dominant processes in our approach does not rely on streamflow but mainly topographic data, what allows predictions of streamflow dynamics in ungauged basins. First, the methods and data of this study are explained. In this section the selection of watersheds, stream flow and explanatory watershed variables are elucidated. In the subsequent sections the results are presented and critically evaluated.

4.2 Methods

4.2.1 Procedures

First, monthly stream flow data periods were selected based on monthly peak flow distributions. Second, collinearities among potential stream variables as well as among watershed variables were identified with correlation analyses. Third, linear regression models were developed for non-collinear dependent variables. Only significant explanatory variables were used (level of

significance $\geq 5\%$). For each stream flow variable, the best model (for which the assumptions for linear regression were met) was selected based on the highest adjusted coefficient of determination (adj. R^2); AIC and BIC were not considered in this study, due to the shortcomings such as the arbitrary relationship between goodness of fit and the amount of parameters (Ward, 2008). For every catchment response variable, all combinations of explanatory variables were tested. Soil and geology data were not used as explanatory regression variables, since detailed soil maps and geological maps were not available in many regions in particular at the required scale for mesoscale watersheds. The developed regressions estimate the explanatory power for catchment response prediction of dominant runoff generation processes per se, and in combination with topographical, landuse, and climate information.

4.2.2 Study Catchments

For this study non-regulated mesoscale watersheds (between 50 km² and 1000 km²) were selected in British Columbia (BC), Canada. We selected stations for which at least 20 years of daily stream flow data were available, and for which the influence of lakes should be minor. According to these criteria, 62 watersheds were chosen, which were distributed across British Columbia with a geographical concentration in the Southeast of BC.

4.2.3 Stream Flow Data

The daily stream flow data were provided by Environment Canada, and the values ranged from 1976 until 2003. The time series overlapped substantially among the 62 gauges. In contrast to Snelder et al. (2005), who focused primarily on ecologically relevant parameters, we assessed measures that quantify catchment response, including specific discharges as well as their rates of increase and decreases as a measure of the runoff dynamics. Table 4.1 explains the different catchment response variables which were assessed and predicted in this study. Dominating storm types and subsequent seasonal hydrograph variability differ substantially across British Columbia. Therefore, not only annual but also monthly stream flow measures were selected. Since we were interested in high flows the 90% quantiles (Q90; for each gauge; quantiles over time) were assessed in this study.

Table 4.1: Catchment response variables (Q90 = 90% quantile).

Category	Variables	Abbr.	Units
runoff	Q90 of the daily specific runoff	QALL	mm
	Q90 of the daily specific runoff of month i	Q _i	mm
runoff increase	Q90 of the daily specific runoff increases within one day	I1ALL	mm/d
	Q90 of the daily specific runoff increases within one day in month i	I1 _i	mm/d
	Q90 of the daily specific runoff increases within five days	I5ALL	mm/5d
	Q90 of the daily specific runoff increases within five days in month i	I5 _i	mm/5d
runoff decrease	Q90 of the daily specific runoff decreases within one day	D1ALL	mm/d
	Q90 of the daily specific runoff decreases within one day in month i	D1 _i	mm/d
	Q90 of the daily specific runoff decreases within five days	D5ALL	mm/5d
	Q90 of the daily specific runoff decreases within five days in month i	D5 _i	mm/5d

4.2.4 Watershed Characteristics

Watershed characteristics were derived from topography, runoff generation, landuse, and climate (Table 4.2). The topographic properties were derived from a 25 m digital elevation model (DEM). The digital elevation model was obtained from different point elevation sources and breaklines for ridges and streams and a hydrological meaningful, seamless DEM for the entire province of British Columbia was calculated (Carver et al., 2009). Digitized stream networks and lake boundaries were available at the scale of 1:20,000 (Ministry of Environment, British Columbia, 2006). Furthermore, road data were provided by the Ministry of Environment, British Columbia, at the same scale.

Dominant runoff generation processes (DRP) areas for channel interception (CHA), Hortonian overland flow (HOF), saturation overland flow (SOF), and shallow subsurface flow (SSF) were derived from the DEM, water bodies, and road data sets. CHA was defined as water which falls into channel grid cells. HOF areas were defined as roads; SOF areas were assumed to be found at locations with either a topographic index higher than 12 or vertical distances to the water table of less than 2 m. The vertical distance to the water table was estimated with RSAGA (Brenning, 2008) which approximates water table depths from stream networks, and water table shapes from ground surface curvatures (Olaya, 2004). In contrast to the depth-to-water index by Murphy et al. (2009), VE was an approximation to derive the vertical distance of a grid cell to a theoretical groundwater table (Etzrodt et al., 2002). The thresholds of the SOF areas were based on field work where we compared soil saturation to topography-based models (Rosin et al., 2010a). SSF areas were delineated where the gradient to the stream along flow paths exceeded 30%. It was assumed that only one process is dominating at one location, and that fast processes dominate

over slower processes (CHA > HOF > SOF > SSF). Furthermore, only DRP areas were considered which were connected to the channel along the flow path (Rosin et al., 2010b). Landuse data (classified into bare surfaces, forests and glacier coverage) were extracted from the Baseline Thematic Mapping data set (Ministry of Sustainable Resource Management, British Columbia, 2001). Mean annual and mean monthly precipitation for each watershed were calculated from the climate normals at a 400 m raster based on data from 1961 until 1990 (Price et al., 2000). The precipitation period precedes the runoff data (1976 until 2003), but the overlap of both periods is substantial.

Table 4.2: Watershed variables (predictor variables of the six regression models).

Category	Variables	Abbr.	Units	Description
topography	area	ARE	km ²	catchment area
	elevation	ELE	m	mean catchment elevation
	slope	SLO	radian	mean catchment slope (local slope; 2 nd degree polynomial by Zevenbergen and Thorne, 1987)
	circularity	CIR	-	ratio of the catchment area to a the area of a hypothetical circular catchment with the same perimeter
runoff generation	channels	CHA	-	approximation of the stream network density: amount of water grid cells per amount of cells per watershed
	Hortonian overland flow	HOF	-	ratio of Hortonian overland flow areas to catchment area
	saturation overland flow	SOF	-	ratio of saturation overland flow areas to catchment area
	shallow sub-surface flow	SSF	-	ratio of shallow subsurface flow areas to catchment area
landuse	forest	FOR	-	ratio of forested areas to catchment area
	glacier	GLA	-	ratio of glaciated areas to catchment area
	bare	BAR	-	ratio of bare areas to catchment area
climate	annual precipitation	PAN	mm/year	mean annual precipitation; normals from 1961 – 1990
	May precipitation	PMA	mm/month	mean May precipitation; normals from 1961 to 1990
	October precipitation	POC	mm/month	mean October precipitation; normals from 1961 to 1990

4.3 Results

4.3.1 Stream Flow

Monthly stream flow data periods were selected based on distributions over years of monthly peak flows (i.e. highest daily flow per month; Figure 4.1), since they reflect preliminary storm signatures: low elevation watersheds (which are mainly located at the coast) are dominated by fall storm systems (October to December; see Figure 4.1) and high elevation watersheds by snow melt (mainly in May; see Figure 4.1). As a result, the flow variables were investigated for the

months of May and October. The month of October was selected instead of November or December - which exhibited higher flows at low elevation - to avoid complex mixtures of snowfall and rainfall processes. In this study high flow values (90% quantiles) were addressed, since the concept of dominant runoff generation processes has been developed for intensive storms (Scherrer and Naef, 2003).

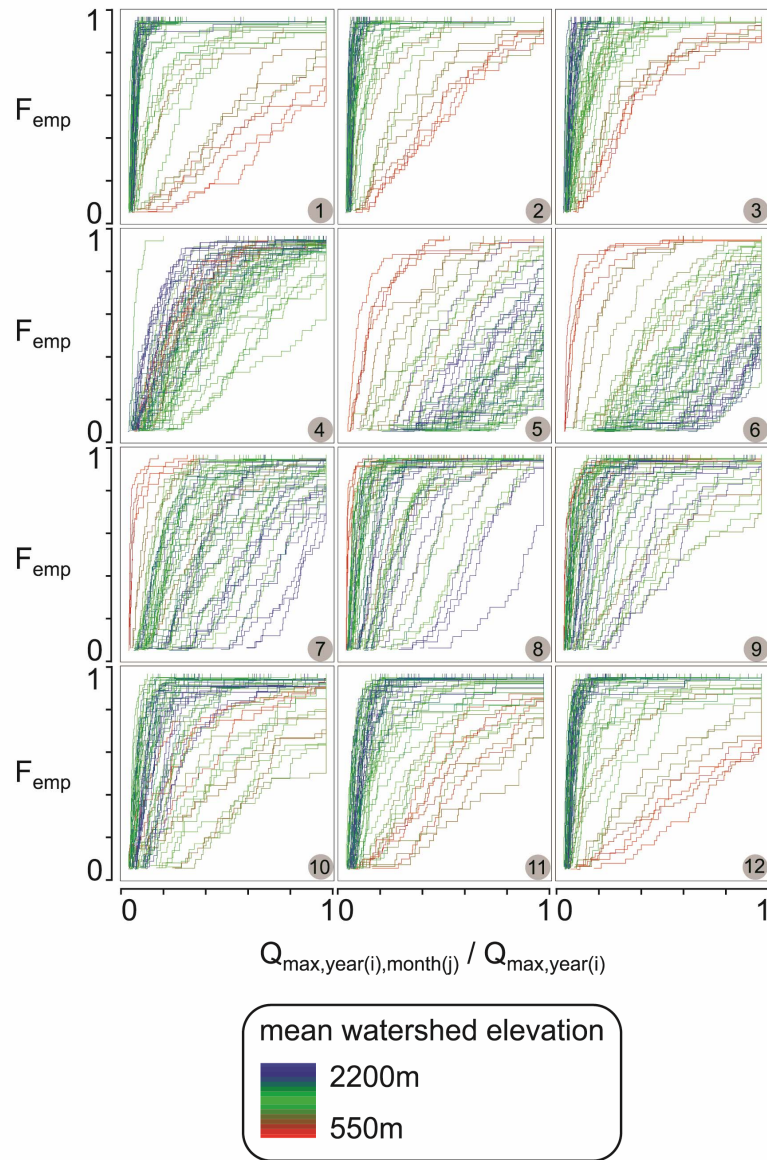


Figure 4.1: Monthly peak flow distributions standardized by annual peak flow are shown for the 62 gauges used in this study. The number in the lower right corner of each panel indicates the month.

Within catchment response variables (Table 4.1) high correlations were found between the 90% quantiles of specific discharge increases and decreases (Figure 4.2). The coefficients of correlation ranged from 0.95 to 0.99. The slope of a linear relationship between increase and decrease varied, in particular between the months May and October; a steeper slope was observed in May than in October as well as for one day (May: 0.87, October 0.47) as over five days (May: 0.86, October: 0.75). In particular for October, approximately a quarter of the assessed catchments exhibited much larger flow increases than the other catchments; despite the differences between these two catchment groups, both groups followed the same regression line. Due to the clear relationship between increase and decrease of the specific discharge, relationships to explanatory catchment variables were only assessed for the flow increase.

High correlation coefficients were discovered between the increase in specific discharge over one day or five days for all assessed time periods (*ALL*: 1.00, *MAY*: 0.98, *OCT*: 1.00). Therefore, models with watershed characteristics were only tested for daily increases in this study. The correlation coefficients between the specific discharge and the increase of the specific discharge were not high, and varied substantially (*ALL*: 0.69, *MAY*: 0.94, *OCT*: 0.98). Therefore, models were tested for specific discharge and its increase at different periods of the year (*ALL*, *MAY*, *OCT*).

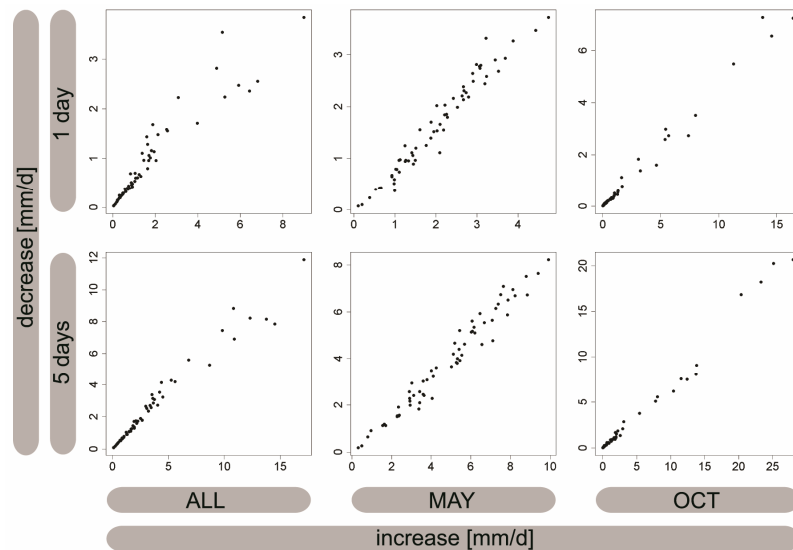


Figure 4.2: Relations between daily stream flow increase and decrease. The relationship between 90% quantiles of specific runoff increase and decrease is shown for the entire year (*ALL*), the months May (*MAY*) and October (*OCT*). The stream flow increase was assessed over periods of one day and five days.

4.3.2 Watershed Characteristics

Moderate correlation coefficients between catchment characteristics were observed (Table 4.2). Most correlation coefficients fell within a range of -0.63 to 0.62. Higher correlation coefficients were observed between the shallow subsurface flow ratio and local slope (0.94), and the annual precipitation to the May (0.77) and October precipitation (0.95). Only slight correlations were observed between dominant runoff generation processes (Table 4.2). The distribution of DRP ratios of the watershed varied differently among processes (Figure 4.3). The HOF areas were limited to less than 5%, CHA to less than 10% of the watershed areas. The SOF and SSF area ratio were considerably larger. In particular, SSF was found to have a high variability across watersheds; the SSF ratios ranged from 0.5 to 52%.

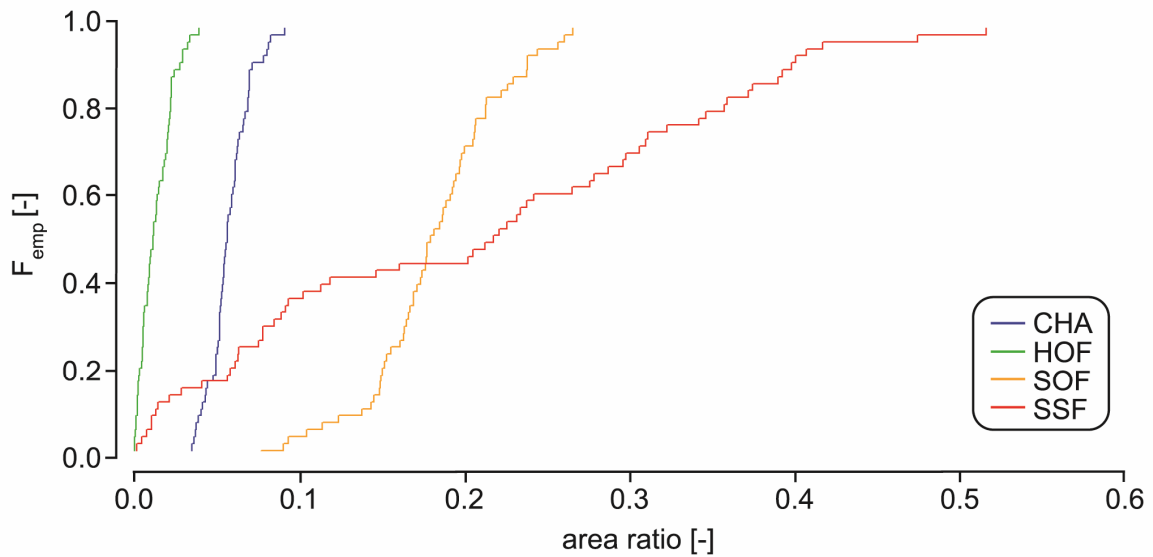


Figure 4.3: DRP proportion in different watersheds. The distributions of DRP areas proportions (e.g. SOF area divided by the watershed area) for different watersheds is displayed.

Table 4.3: Correlation coefficients between catchment characteristics. Variables descriptions can be found in Table 4.2.

	Topography				DRP			Landuse			Climate		
	ELE	SLO	CIR	CHA	HOF	SOF	SSF	FOR	GLA	BAR	PAN	PMA	POC
ARE	0.05	0.32	-0.43	0.25	-0.22	0.17	0.35	-0.33	0.19	0.17	0.22	0.11	0.21
ELE		0.33	0.21	-0.02	-0.29	0.22	0.21	-0.32	0.20	0.40	-0.39	0.07	-0.45
SLO			0.09	0.23	-0.48	0.11	0.94	-0.61	0.44	0.41	0.52	0.60	0.44
CIR				0.03	-0.12	0.01	0.11	-0.09	0.21	-0.01	-0.07	0.05	-0.02
CHA					-0.41	0.46	0.36	-0.29	0.09	-0.05	0.17	0.06	0.24
HOF						-0.15	-0.51	0.22	-0.36	-0.20	-0.09	-0.01	-0.16
SOF							0.10	-0.51	0.16	0.07	0.08	0.06	0.11
SSF								-0.53	0.44	0.35	0.54	0.53	0.50
FOR									-0.63	-0.34	-0.27	-0.24	-0.28
GLA										0.23	0.23	0.07	0.30
BAR											0.03	0.16	-0.05
PAN												0.77	0.95
PMA													0.62

4.3.3 Regression Models

The value of dominant runoff generation processes information on the prediction of catchment response varied among stream flow measures and for different degrees of watershed data. In Figure 4.4 the adjusted coefficients of determination across watersheds are presented for the best regression models based on catchment data. Dominant runoff generation areas per se had high explanatory power for QALL, QMAY, and I1MAY. However, DRP areas could not explain much of the variability in QOCT, I1ALL, and I1OCT. For these measures, the prediction was substantially better using topographical data rather than DRP. However, the prediction of QOCT, I1ALL, and I1OCT did clearly improve when a combination of topographical and DRP data was applied. Independent of the incorporation of DRP, landuse did not improve the prediction. In contrast to landuse, precipitation data has a substantial impact on catchment response predictions. However, the DRP impact was marginal if topography, landuse, and climate data were available. Generally, the highest benefit of DRP data was found for dependent variables with low adjusted coefficient of determination.

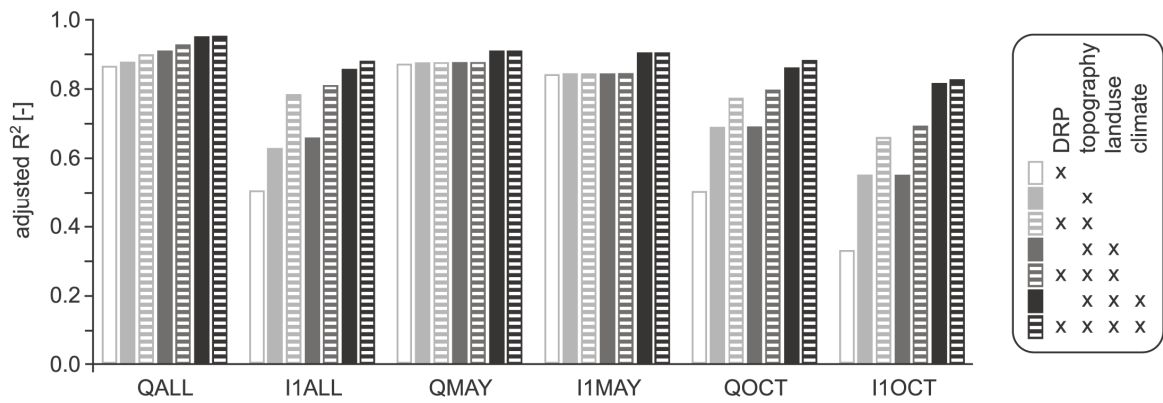


Figure 4.4: Explanatory power of DRP concepts. For DRP data per se and in combination with topography, landuse, and climate data the adjusted coefficient of determination (adjusted R^2) across catchments are shown for different stream flow measures (QALL, I1ALL, QMAY, I1MAY, QOCT, I1OCT).

The best prediction model structures are presented in Table 4.4. The three most successful models considering only DRP as predictor values (QALL, QMAY, I1MAY) consisted of the variables SOF and SSF sometimes combined with HOF. However, the less successful DRP models (QOCT, I1ALL, I1 OCT) did not contain the variable SOF. SOF was often found when dominant runoff generation processes were combined with topographic, landuse or climate variables; conversely, SSF was often replaced by the local slope in combined models. CHA was only used in one model.

The local slope and the mean watershed elevation were the most used topographic variables. The regression coefficients of ELE were negative. When climatic information was available, elevation was substituted by precipitation variables. Predictions of QMAY and I1MAY were dominated by SLO if no climate information was accessible. In contrast to bare areas, the proportions of glaciated or forested areas were often used in the regression models. However, landuse information was not incorporated in models for predicting stream flow increase, if climate data were available. With climate information, all predictions were strongly explained by precipitation. Monthly precipitation was more important than annual precipitation for monthly stream flow predictions.

Table 4.4: Model structures with best predictions.

dependent variables	independent variables				adj R ²	best models
	DRP	topography	landuse	climate		
QALL	X				0,86	15.23*SOF+17.76*SSF
QALL		X			0,88	27.89*SLO-0.0022*ELE
QALL	X	X			0,90	22.55*SOF+25.49*SLO-0.0042*ELE
QALL		X	X		0,91	0.0027*ARE+22.31*SLO-0.003*ELE+2.939*FOR+23.32*GLA
QALL	X	X	X		0,93	61.12*HOF+20.16*SOF+22.67*SLO-0.0041*ELE+21.24*GLA
QALL		X	X	X	0,95	7.479*SLO-2.936*CIR+20.62*GLA+0.0033*PAN
QALL	X	X	X	X	0,95	9.646*SOF+6.048*SSF-3.476*CIR+20.11*GLA+0.0032*PAN
QMay	X				0,87	101.04*HOF+22.91*SOF+19.88*SSF
QMay		X			0,88	26.61*SLO
QMay	X	X			0,88	26.61*SLO
QMay		X	X		0,88	26.61*SLO
QMay	X	X	X		0,88	26.61*SLO
QMay		X	X	X	0,91	12.65*GLA+0.1146*PMA
QMay	X	X	X	X	0,91	7.345*SSF+0.0998*PMA
QOCT	X				0,50	14.81*SSF
QOCT		X			0,69	0.0037*ARE+21.52*SLO-0.0065*ELE+10.95*CIR
QOCT	X	X			0,77	38.89*SOF+22.82*SLO-0.0075*ELE
QOCT		X	X		0,69	0.0037*ARE+21.52*SLO-0.0065*ELE+10.95*CIR
QOCT	X	X	X		0,80	35.55*SOF+21.56*SLO-0.0083*ELE+3.357*FOR+13.70*GLA
QOCT		X	X	X	0,86	-0.0008*ELE+0.0322*POC
QOCT	X	X	X	X	0,88	17.01*SOF+7.484*SLO-0.0036*ELE+0.0233*POC
I1ALL	X				0,50	50.58*HOF+5.180*SSF
I1ALL		X			0,63	0.002*ARE+8.437*SLO-0.0028*ELE+5.424*CIR
I1ALL	X	X			0,78	53.43*HOF+17.15*SOF+10.24*SLO-0.0036*ELE
I1ALL		X	X		0,65	12.05*SLO-0.0027*ELE+2.717*FOR
I1ALL	X	X	X		0,81	42.27*HOF+15.81*SOF+9.752*SLO-0.0041*ELE+1.909*FOR+5.392*GLA
I1ALL		X	X	X	0,86	-0.0009*ELE+0.0023*PAN
I1ALL	X	X	X	X	0,88	8.997*SOF-0.0017*ELE+0.002*PAN
I1MAY	X				0,84	6.735*SOF+3.924*SSF
I1MAY		X			0,84	5.739*SLO
I1MAY	X	X			0,84	5.739*SLO
I1MAY		X	X		0,84	5.739*SLO
I1MAY	X	X	X		0,84	5.739*SLO
I1MAY		X	X	X	0,90	0.0005*PAN+0.017*PMA
I1MAY	X	X	X	X	0,90	0.0005*PAN+0.017*PMA
I1OCT	X				0,33	9.802*SSF
I1OCT		X			0,55	19.61*SLO-0.0056*ELE+8.953*CIR
I1OCT	X	X			0,66	34.49*SOF+18.13*SLO-0.0067*ELE
I1OCT		X	X		0,55	19.61*SLO-0.0056*ELE+8.953*CIR
I1OCT	X	X	X		0,69	30.70*SOF+17.41*SLO-0.0076*ELE+3.535*FOR+11.31*GLA
I1OCT		X	X	X	0,82	-0.0012*ELE+0.028*POC
I1OCT	X	X	X	X	0,83	33.45*CHA-0.0021*ELE+0.0252*POC

4.4 Discussion

The purpose of this study was to investigate the influence of dominant runoff generation processes on stream flow predictions given different types of catchment information. Data limitation is a relevant factor, because runoff simulations cannot always rely on detailed precipitation and stream flow information (e.g. Corral et al., 2000). For each level of data availability stream flow predictions should concentrate on dominating factors (Wagener et al., 2008). That is why both the delineation of dominant runoff generation processes and the level of data availability were hierarchically assessed in this study.

4.4.1 Catchment Response

The high correlation between stream flow increase and decrease indicates coherence between catchment functions. The combination of sharp runoff rise with sudden decline could be suggestive of a relationship among partition toward fast processes, limited storage, and/or immediate release. On the other hand, precipitation has been described to dominate stream flow dynamics (e.g. Nicotina et al., 2008). In particular the relation of storm intensity, type, volume, and position relative to contributing areas in a watershed determined catchment functions (Black, 1997). Consequently, the substantial precipitation differences in British Columbia dominate the catchment response signatures, which led to high correlations of stream flow increase and decrease among watersheds. That was why variability in the 90% quantiles of annual stream flow originated from different periods of the year for different watersheds in British Columbia. In contrast to annual runoff measures, May and October stream flow variables provided a better basis for process-based interpretation.

4.4.2 Watershed Characteristics

In general, the distribution of active areas in a watershed might have an impact on the dynamics of the catchment response. McGlynn and McDonnell (2003) assessed the ratio of riparian to hillslope zones to investigate catchment runoff. In our study, the concept of McGlynn and McDonnell (2003) was extended to several runoff generation processes. Similar to Snelder and Biggs (2002) the stream network played a prominent role in our hierarchical process delineation. Even though the dominant runoff generation processes were delineated from different, mainly topographical information, some correlations in Table 4.3 between watershed characteristics need to be discussed. The high correlation between the mean local slope and the SSF area

proportion – which were derived from the average gradient along the flow path – was fairly surprising. It could be explained by the variability of the gradient along flow paths within, and great variability between watersheds.

Furthermore, the correlation between CHA and SOF was lower than expected. CHA could be seen as an approximation of the stream network density. However, in the assessed watersheds a higher stream network density did not imply more saturation areas. This is in agreement with a SOF-model evaluation with field data, which indicated more saturation areas due to high topographic indices than low distances to groundwater table (Rosin et al., 2010a). The high negative HOF/SSF correlation originated from the fact that fewer roads could be found in steep terrain.

A negative correlation was found between elevation and precipitation, because the mean annual precipitation in British Columbia clearly declines from the low elevation coast to the higher elevation interior. At the coast, the annual precipitation was dominated by intensive fall storms. We hypothesized that this might explain why the mean annual precipitation was better correlated to October than to May precipitation.

4.4.3 Regression Models

The regression models are discussed under consideration of water input processes such as snow melt and rain storms in British Columbia. QMAY and IIMAY were dictated by snow melt processes, QOCT and IOCT by rain storms. However, the process influences were more complex for the annual measures QALL and IALL. Therefore, we studied the time of high discharge (=discharge exceeding the 90% discharge quantile) for each gauge. Analogously, the time of high discharge increases (=discharge increases exceeding the 90% quantile of the discharge increase) was assessed for each watershed. We found a small influence of fall storms (months September to December) for QALL: only 30.6% of high discharge occurred in fall (90% quantile across watersheds). Analogously, 50.6% of high discharge increases were found in fall. Consequently, IALL was more influenced by fall storms than QALL. The adjusted coefficients of determination (adj. R^2) of the regression models for QALL (which ranged from 86% - 95%) were comparable to a study by Roland and Stuckey (2008), in which elevation and landuse variables explained 88% - 96% of the variability of streamflow (90% quantiles). Vogel et al. (1999) explained the variability in mean and standard deviation of annual runoff with regression

models based on climate, geomorphic, and hydrologic variables. The adjusted coefficients of determination by Vogel et al. (1999) ranged from 90.2% to 99.8%, and were clearly higher than in our study (adj. R^2 fall between 33% and 95%; Figure 4.4).

The worst topography-based runoff prediction was found for the rain storm dominated measures QOCT, IIOCT, and IIALL (Figure 4.4). With topography, neither the volume nor dynamics of the catchment response were predicted reasonably. Considering dominant runoff generation processes resulted in a substantial improvement. The improvement could have occurred because the conditions of intense precipitation agreed best with assumptions of the dominant runoff generation processes concept. Scherrer and Naef (2003) demonstrated that dominant runoff generation processes have value for intensive precipitation events. For intense rainfall, not only overland flow but also subsurface flow might contribute to rapid discharge increases (Beven, 2001). Due to the high correlation between SSF and the local slope, subsurface flow generation areas were only implicitly considered in the regression models.

On the other hand, DRP did not improve the topographic predictions of snowmelt dominated measures (QMAY, IIMAY). During snowmelt, runoff generation is often limited by the driving gradient to the channel and spatial saturation patterns. Consequently, the local slope of a watershed exhibited high explanatory power.

In reality, landuse has a major impact on runoff formation due to its influence on interception, evapotranspiration, or its effects on snowmelt, infiltration, and subsurface flow. In general, landuse information is important in runoff generation models (e.g. Romanowicz et al., 2005). We conjectured that land use was not an important factor in the regression analysis because most watersheds were forested, and hence between watershed variability was low.

Dominant runoff generation processes can be used to predict the partitioning of water input in a catchment. However, DRP need to be combined with water input variables for catchment response predictions. In this study, reasonable runoff prediction was achieved due to the inverse relationship between precipitation and elevation. If climate data are available, the watersheds could initially be classified by precipitation regime (e.g. Ali et al., 2010); in a second step, typical runoff response could be estimated with the concept of dominant runoff generation processes. It should be noted that the value of soil data was not tested in this study, even though

it could be as important as climate information (Zehe et al., 2005). Furthermore, only monthly precipitation but not precipitation intensity could be incorporated in this study. Precipitation intensity and soil property data could improve the catchment response predictions by realistic representations of antecedent soil moisture (Zehe and Bloeschl, 2004).

4.5 Conclusions

In this study we assessed the usefulness the dominant runoff generation process concept for catchment response prediction for different levels of data availability. We found a considerable potential of DRPs to enhance predictions of rain storm influenced catchment response measures (e.g. October specific discharge, increase of October specific discharge) if only topographical information was available. However, dominant runoff generation processes per se had little explanatory power for predictions of catchment responses to rain storms. The effect of dominant runoff generation process on stream flow predictions was fairly small in combination with comprehensive data sets (topographic, landuse, and climate data). On the whole, DRP approaches exhibited major benefit for predictions in data-scarce watersheds. For future research we recommend evaluating contributing source areas of catchments for example with isotope analyses (Laudon et al., 2007). Such contributing area evaluations could help to establish dominant runoff generation process area delineations as a spatially distributed prediction support concept for catchment response simulations ungauged basins.

4.6 Acknowledgements

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5 LARGE SCALE ASSESSMENT OF LAND COVER CHANGE IMPACTS ON PEAK FLOW IN UNGAUGED BASINS¹

5.1 Introduction

The assessment of the impact of land cover or land use change on the hydrological cycle is still a challenging task and reliable answers, for example, of the impact of the still rapidly progressing changes to tropical rain forests on the hydrological cycle, are still missing (Achard et al., 2002; Marengo et al., 1994). In North America in recent years, wildfires and insect infestations have caused rapid changes in the land cover. For example, the increase of large scale forest fires in the Western USA has not been assessed in detail for its hydrological impacts (Miller et al., 2003).

In British Columbia (BC, Canada), the mountain pine beetle (MPB, *Dendroctonus ponderosae* Hopk.) epidemic has caused substantial tree mortality since 1999 (Carroll et al., 2006), particularly of lodgepole pine (*Pinus contorta* var. *latifolia*; Taylor et al., 2006), which is a primary host in British Columbia (Aukema et al., 2006). At the end of 2006, the cumulative infested area exceeded 130,000 km² (Kurz et al., 2008). 78% of the lodgepole pine is expected to be killed until the year 2018. Beyond the vast areas in British Columbia, the infestation is also progressing rapidly into the American Rocky Mountains and into Alberta. In response to the MPB infestation, the forest industry is salvaging as much timber as possible before it becomes unusable. Consequently, in some watersheds more than 50% of the forested area may be logged over the next years, which is significantly higher than historical logging rates in BC. Current regulations do not specify which areas in a watershed should not be logged except for riparian corridors and old growth reserves. The government of British Columbia has been developing tools to assess the MPB and salvage logging impact on peak flow, low flow, bed load transport, suspended sediment and stream temperature for the entire province (Carver et al., 2007).

Most assessment strategies either rely on very simplified models that use empirically derived relations between land cover change and hydrological variables, or hydrological rainfall-runoff models are applied and hydrographs are simulated assuming different land cover scenarios. Both approaches have significant disadvantages for assessing large-scale changes. The empirical

¹ A version of this chapter will be submitted for publication. Rosin, K., Weiler, M., 2010. Large scale assessment of land cover changes on peak flow in ungauged basins.

models are often developed from paired-watershed experiments that studied the effect of land cover on the hydrological response. In smaller scale agricultural studies the differences in overland flow generation were assessed. Larger scale experiments - analyzing the differences of watershed runoff - were mostly initiated to study the influence of forest management and logging on annual runoff and peak flow (e.g. Bosch and Hewlett, 1982; Moore and Wondzell, 2005; Stednick, 1996). Since paired-watershed studies cannot eliminate natural variability, the results are specific to the observed climate, topography, soils and geology. The other strategy to assess land cover changes is to use spatially explicit hydrological models that simulate the small scale processes at the soil-vegetation-atmosphere interface and the large scale runoff generation processes. The models are often detailed physically-based conceptualizations of the hydrological cycle (e.g. DHSVM, SWAT, WASIM-ETH, VIC). Their ability to simulate changes can be satisfactory, but they are time consuming to set up, the watershed area is limited by the chosen grid-cell resolution and computing time, and most importantly, they still need to be calibrated to existing stream flow data (Niehoff et al., 2002; Storck et al., 1998; Van Shaar et al., 2002, Schnorbus et al. 2010). Therefore, their suitability for application to ungauged watersheds is limited and they are impracticable for large areas where small scale changes have to be assessed.

Only a small number of streamflow gauges and climate stations are available in the large mountain pine beetle infested area in British Columbia. There is, therefore, an urgent need for methodologies that are applicable to ungauged basins. Peak flow and effects of land use change in ungauged watersheds has often been estimated by model parameter regionalization (e.g. Hunscha and Bárdossy, 2004). Yadav et al. (2007) introduced a method for regionalizing model independent stream flow indices instead of model parameters to constrain ensemble predictions in ungauged basins. However, most studies in ungauged watersheds have relied on fairly complex hydrologic models (e.g. Storck et al., 1998). Consequently, Sivapalan et al. (2003) called for models that consider the dominant processes for the measures of interest (e.g. peak flow), and suggested incorporating adequate climatic input. Croke and Jakeman (2001) pointed out the benefits of conceptual, distributed hydrologic models to assess land use changes on stream flow.

In this study, we introduced and tested a new approach to assess large scale land cover change impacts on peak flows, which are a major concern in relation to flood hazard, sediment transport and erosion. The approach focused on:

- Explicit consideration of the spatial linkages of land cover changes to hydrological processes at small scales of less than one hectare.
- Applicability to ungauged watersheds through independence of parameters from climatic, pedological and geological properties.
- The prediction of long term average change in runoff conditions rather than individual runoff events.

First, the general concept of the assessment approach is introduced, and the model framework and model components explained. Second, the results of the model application to two regions in British Columbia are evaluated.

5.2 Methods

5.2.1 General Concept

The general philosophy of the proposed concept was that areas which generated more runoff in a watershed during a rainfall or snowmelt event were classified as more sensitive to land cover modification. Hydrological changes in these areas were assumed to cause relatively large change at the watershed scale. This idea dated back to historical findings by Betson (1964), Dunne and Black (1970), and Weyman (1970) insofar as runoff could be generated by multiple processes (variable source area concept), but that areas where these processes took place did not overlap. Betson (1964) demonstrated that contributing areas were almost constant during heavy rainfalls. Dunne and Black (1970) extended Betson's concept to saturation excess overland flow and Weyman (1970) to subsurface flow. Scherrer and Naef (2003) developed a decision tree to identify different dominant runoff generation processes at the plot scale. Schmockler-Fackel et al. (2007) introduced a procedure to identify areas of different generating processes within a GIS framework. Other authors developed similar approaches using different procedures and GIS products, but focusing on the idea that runoff generation areas can be used to predict the response characteristics of watersheds (Tetzlaff et al., 2007; Uhlenbrook et al., 2004; Walter et al., 2000). As advocated by McDonnell (2003), we also believe using first-order runoff generation processes knowledge at the basin scale is an appropriate trade-off between experimental process knowledge and model complexity.

The purpose of the approach presented here is to classify the sensitivity of peak flow regimes in third-order watersheds (less than 200km²) to land cover modification due to MPB over a large area. To guarantee applicability at the large scale in ungauged catchments, this classification is solely based upon spatial information of

- a) Input characteristics derived from monthly gridded maps of climate normals;
- b) Runoff generation processes derived from GIS data available for the entire province of British Columbia.

Figure 5.1 presents the general methodology of this study. Areas in a watershed that are sensitive in terms of changes in peak flow in the main river channel are derived from the following model components:

- 1) Input component: mapped peak-flow-generating hydro-climate input for each defined watershed.
- 2) Runoff generation component: delineated dominant peak-flow-producing hydrologic processes at the watershed scale.
- 3) Land cover modification component: modified input depending on the vegetation cover according to stand level research.

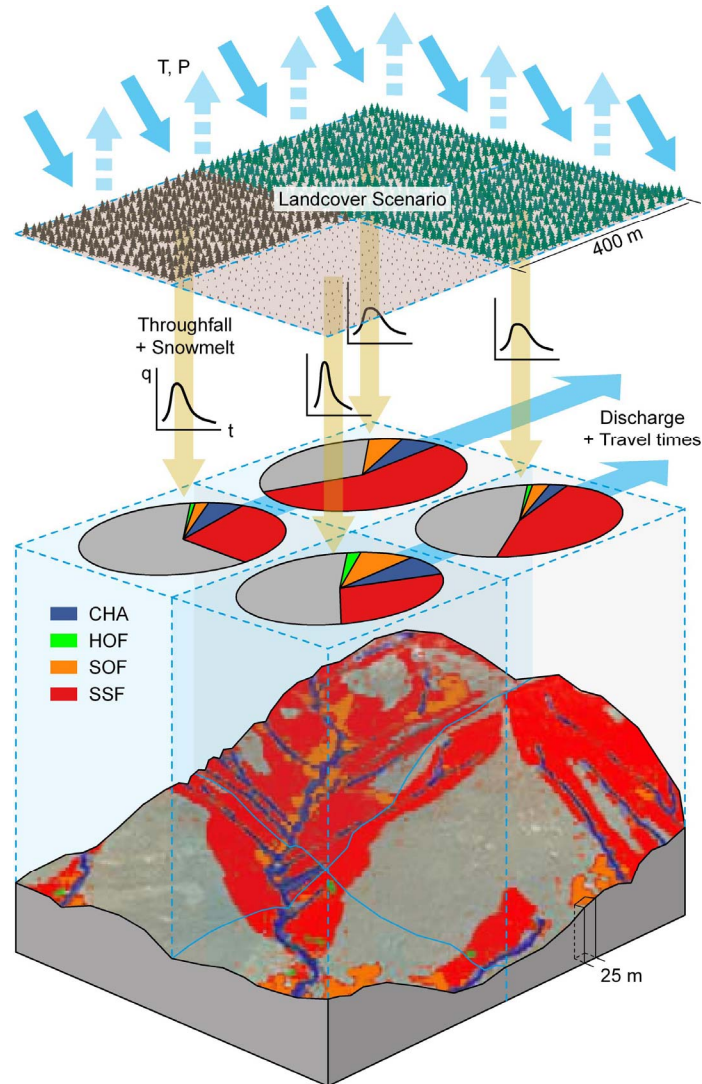


Figure 5.1: Model structure. The interaction of the input component (hydro-climate input for each watershed), runoff generation component (peak flow based on dominant runoff generation process areas), and land cover modification component are shown. Abbreviations: T = temperature, P = precipitation, CHA = channel interception, HOF = Hortonian overland flow, SOF = saturation excess overland flow, SSF = shallow subsurface flow.

5.2.2 Input Component

The peak flow regime of a watershed is related to its precipitation regime (snowmelt-dominated, rainfall-dominated, and transitional). In a snowmelt-dominated watershed, peak flow is usually initiated by snowmelt during the spring freshet. The input in snowmelt-dominated watersheds is spatially and temporally highly variable with early melt in the lower portion and south-facing slopes of the watershed, and late melt in the higher parts and north-facing slopes of the basin (Jost et al., 2007). Hence, only certain areas in the watershed produce runoff during peak flow. The input in rainfall-dominated watersheds depends mainly on elevation.

The input model uses the mean monthly climatic precipitation and temperature data that were derived using the PRISM methodology for a 400 m grid spacing for the province of BC (Spittlehouse, 2006). The input model uses mean daily temperature and precipitation at a site that is interpolated from monthly climatic data to define the rate of snow accumulation and snow melt. Precipitation falls as snow if air temperature $T < T_0$ and as rain otherwise. Snow melts according to the degree-day factor K (mm/day/°C). The threshold air temperature T_0 is set to 0 °C and K to 3.0 mm/day/°C (Kuusisto, 1980; Rango and Martinec, 1995). The degree day factor depends on many factors, including the relation of short to long wave radiation, elevation, and topography (Rango and Martinec, 1995).

The input model calculates SWE and hence snow melt for each grid cell for every day for one hydrological year starting in September and ending in August. The model extracts all grids within each third order watershed in BC and calculates the average melt time series for each defined watershed (see example in Figure 5.2, left).

The average melt flux time series for each watershed is then used to determine the date of the maximum melt. For this date, the snowmelt flux of each individual grid is extracted and stored as the peak melt flux for each watershed (Figure 5.2, right). As to be expected from this watershed with an elevation range of more than 1000 m, the lower portion of the watershed was already melted out whereas the maximum melt occurred in the upper half of the watershed.

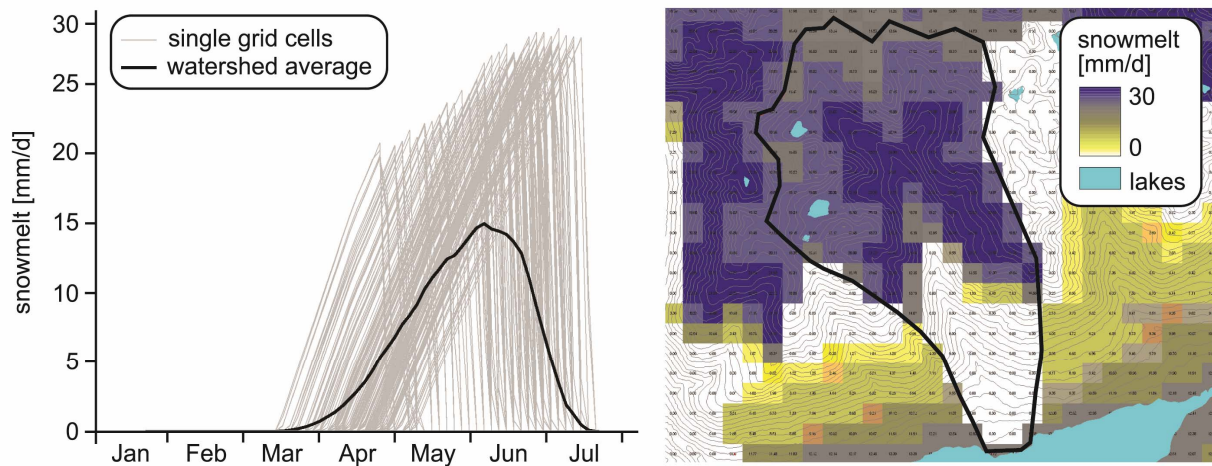


Figure 5.2: Climatic snowmelt response in Redfish Creek. The watershed is located near Nelson (Southeast British Columbia) and has catchment size of 28.6 km². Left: snowmelt time series for each grid cell and the watershed average. Right: Peak snowmelt flux on June 2.

5.2.3 Runoff Generation Component

Four different runoff generation processes that can contribute to stream flow are mapped: channel interception (CHA), Hortonian Overland Flow (HOF), Saturation Overland Flow (SOF) and Shallow Subsurface Flow (SSF). The location of these dominant runoff processes (DRP) areas in a watershed is related to a combination of factors such as relief, slope, aspect, soil properties, drainage density, drainage pattern, and hillslope curvature. The mapping procedure is based on a 25 m grid size resolution, implemented into the SAGA software and is described briefly for each DRP.

5.2.3.1 Channel Interception

Hewlett and Hibbert (1963) defined channel interception (CHA) as the process that collects water into a stream that falls directly from clouds or indirectly from vegetation onto the stream surface or the saturated portions of the riparian zone. Channel interception is defined for all grid cells that are intersected by a stream and therefore include the channel and part of the riparian zone.

5.2.3.2 Hortonian infiltration excess overland flow

Kirkby (1969) stated that Hortonian infiltration excess overland flow (HOF) can be understood as ‘the flow which occurs when rainfall intensity is so large that not all the water can infiltrate’. Cappus (1960) defined infiltration excess areas as roads, compacted soils, and plastered paths.

The model defines roads and areas with low infiltration capacity - e.g. regions with recent fire history - as HOF areas if there is connectivity to the stream network. Connectivity is calculated from the horizontal overland flow distance to the stream and should not exceed 300 m.

5.2.3.3 Saturation Overland Flow

Due to topographic features, some zones of a catchment are much more susceptible to saturation and subsequent saturation overland flow (SOF). Kirkby (1969) identified such areas as those adjacent to perennial streams, slopes with concave profile, hollows and hillslopes with shallow soil. The topographic wetness index has been developed and tested to delineate saturated concavities and topographic hollows if lateral flow above an impermeable bedrock layer occurs (e.g. Guentner et al., 1999). We use a version that is based on a modified catchment area calculation (Boehner et al., 2002) and replaced the local slope with the slope to the down slope stream segment (Merot et al., 2003). Areas with a wetness index larger than 10 and underlying low permeable bedrock are mapped as SOF areas. In addition, riparian zones and areas close to a water body become frequently saturated since the groundwater table is close to the soil surface and the moisture deficit is low (McGlynn and Seibert, 2003). Arp (2005) developed a methodology to map these areas by iteratively interpolating the elevation of all open water areas (lakes and streams). This approach was implemented into SAGA and is used to calculate the vertical distance of the groundwater table to the soil surface for all grid cells. In contrast to the depth-to-water index by Murphy et al. (2009), vertical distance of the groundwater table in this study was an approximation to derive the vertical distance of a grid cell to a theoretical groundwater table (Etzrodt et al., 2002). We assume areas to be saturated during peak flow if the vertical distance is less than 2 m (Rosin et al., 2010).

Shallow Subsurface Flow (SSF): Whipkey (1965), Hewlett and Hibbert (1963) and others demonstrated that subsurface flow is an important process for its contribution to fast catchment responses after rain storms or snowmelt events for areas with an impeding layer in the soil. Hewlett and Hibbert (1963) claimed that in most well-vegetated, humid watersheds, subsurface flow is predominant for various storm types. In British Columbia, soils covered with forests are often shallow and are characterized by impeding layers (either fine textured moraine or bedrock). In the proposed framework, steep slopes with a relative short distance to the channel are defined as SSF areas, given they are underlain with an impermeable layer of soil or bedrock. With regard

to steepness, the average slope to the stream – and not the local gradient – is of interest. The model defined SSF areas as those with average gradients along flow paths steeper than 30%.

The runoff generation model component combines the mapped DRP areas and generates a DRP map that shows the processes for a watershed or a larger area of interest. DRPs are mapped hierarchically: $CHA > HOF > SOF > SSF$. Areas that are not mapped as one of the four processes are considered not to contribute to peak flow. Figure 5.3 shows an example of mapped DRPs.

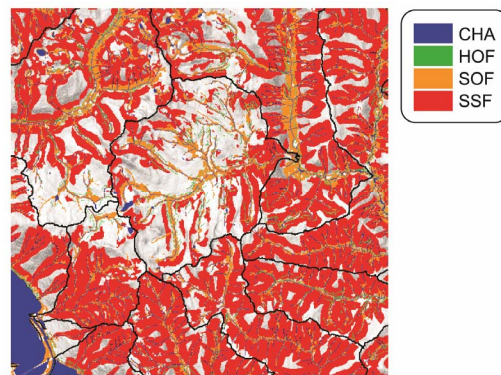


Figure 5.3: Predicted DRP areas. Predictions of dominant runoff generation process areas are shown for several watersheds within the Kooteney Lake TSA.

5.2.4 Land Cover Modification Component

In this component, the actual runoff from each grid cell for a given scenario (see result section) is calculated based on the actual input due to vegetation modification and contribution from each runoff generation process. Data for the vegetation modification originated from various studies at the stand level scale analyzing the influence of vegetation, in particular forests, on rainfall and snowmelt (Table 5.1). In this study, the input modification is presented for snowmelt conditions. Table 5.1 lists several stand level studies, mostly in the Pacific Northwest, studying the difference of snow melt between forests and open land. Only a few studies examined the melt rate difference on a short time scale (e.g. daily) and even fewer studies rely on a larger number of samples to establish more general relations between forest and open land. When focusing on the studies undertaken using larger data sets and in forests that are similar to BC, a reduction of approximately 50% in snow melt in the forest compared to an open is reasonable to assume (Table 5.1).

Table 5.1: Snowmelt rates. Selected stand level studies comparing snow melt rates between forested and open land.

Reference	Melt rate [mm/d]		Forest/Open (%)	Description
	Forest	Open		
Hardy and Hansen-Bristow (1990)	5.8	9.8	59	Average seasonal snow melt rates; Montana.
Jost et al. (2007)	4.1	6.1	67	Multiple regression analysis of average melt rate (20 days in April) including elevation, aspect and forest cover (lodgepole pine)
Kittredge (1953)	7-19	12-24*	48-58	Regression analysis of daily melt rates in different forest stands (white fir, ponderosa pine, mixed conifer) against snowmelt in open over five freshet seasons
Teti (2007)	3-4	5-6.5	50-65	Average melt rates in spring 2007 in lodgepole pine forest in Central BC
Toews and Gluns (1986)	8	11	73	West Kootenay Area, in the South of British Columbia; average seasonal melt rates.
Whitaker and Sugiyama (2005)	6.1-7	12.3	49-57	Lysimeter study of average daily melt rates in a larch and cedar forest in Japan
Winkler et al. (2005)	3	8	38	Measured average melt rate (snow tube and lysimeter in spruce-fir pine stands with different characteristics in Southern British Columbia).

Since detailed GIS data about forest characteristics were available for the study sites, we decided to include canopy closure as a second variable to differentiate snow melt among different stands. Several studies show that canopy closure (or crown coverage), which is defined as a proportion of the ground area covered by the vertical projection of the tree crowns, is linearly related to differences of snow accumulation and melt rates (D'Eon, 2004; Kittredge, 1953; Winkler and Roach, 2005). The main reason for this relationship is the approximately linear relation between canopy closure and net solar radiation below the canopy (Hendrick et al., 1971; Tarboton and Luce, 1996). In the model, we used a linear relation between canopy closure and snowmelt assuming a maximum reduction of 50% at a canopy closure of 80% (Winkler and Roach, 2005).

Additionally, the effect of MPB infestation on the input modification was considered. Since MPB only kills pine trees, we include the percentage of pine coverage to estimate the maximum proportion of trees that can be killed within a stand. Research at the stand level studying the impact of dead trees on snow accumulation and melt are just underway. After the MPB attacked the trees, the needles first become red (red attack) and after a year the needles fall off and only the tree boles and branches will remain for several years (gray attack). The parameterizations for MPB are based on the gray attack stage since this is the more persistent condition (Rex and Dube, 2006). Initial results from several studies in MPB infested stands revealed that gray attack

stands are closer to a healthy forest than a clear cut in respect to snow accumulation and ablation (Boon, 2007; Teti, 2007). A study comparing larch, cedar and open sites in Japan – a leafless larch forest should be comparable to a gray attack pine stand – showed that snow melt at the larch site was even lower than at the denser cedar site (Suzuki and Ohta, 2003). Rowland and Moore (1992) observed 20 to 50% lower maximum daily solar irradiances under a leafless deciduous forest compared to above canopy solar irradiance. Since the research about the influence of MPB attacked stands is not definitive yet, we conservatively parameterized the snow melt rate in gray stands to be half in-between a healthy stand and a clear cut.

The contribution from each runoff generation process area is defined based on the process understanding and its response during peak melt rate input. We define a runoff contributing factor, RC , for each process area that is multiplied with the modified input to simulate a peak flow contribution (mm/day) for each grid cell. The factors are: channel interception: $RC = 1.0$; Hortonian overland flow: $RC = 0.9$; saturation overland flow $RC = 1.0$; subsurface flow: $RC = 0.7$; no dominate runoff generation process defined: $RC = 0.0$. The average daily peak flow (m^3/s) for each sub-watershed is calculated by multiplying the watershed area with the average peak flow contribution of the watershed.

5.2.5 Study Area

The model was applied to two timber supply areas (TSA) in British Columbia, the Kooteney Lake and Prince George TSA (Table 5.2). Of the Prince George TSA only the sub-region Block C was assessed; for reasons of simplicity, the assessed area was still named Prince George TSA in this study. Physiography and geology of the two areas varied considerable; therefore, they provided an adequate test bed for our introduced approach.

Table 5.2: Site characteristics.

	Kooteney Lake TSA	Prince George TSA
size	12,000 km ²	75,000 km ²
location	Southeast of British Columbia	North central interior of British Columbia
number of snow sites	25	9
elevation range of snow sites	650 m – 2230 m	690 m – 1620 m
number of watersheds with observed peak flow	13	2

5.2.6 Prediction Evaluation

The accuracy of the predicted daily peak flow depended on the accuracy of the simulated snowmelt fluxes according to the input component, the DRP map, and the assigned runoff contributing factors. The correctness of these factors was tested stepwise.

First, snowmelt fluxes and input component were evaluated. The agreement between simulated snow water equivalent (SWE) time series and historical snow course data of 34 snow sites was quantified with the mean error ME (mean difference of observed minus simulated SWE), the root mean square error RMSE, and the coefficient of determination (R^2). Snowmelt was continuously simulated for the values at the location of the snow courses were extracted for the first day of April, May, and June, and evaluated with manual snow survey data from the Water Stewardship Division of the BC Ministry of the Environment (Ministry of Environment, British Columbia, 2009). Only sites with at least ten years of data were selected to calculate monthly average SWE at each site.

Second, the predicted daily peak flow was compared to stream flow from Water Survey of Canada (WSC) gauging stations in and close to the two TSAs with at least five years of daily stream flow records. For the Kootenay Lake TSA, 13 sites with drainage areas between 26 km² and 282 km² were selected; two sites with drainage areas of 414 and 881 km² were chosen for the Prince George TSA (Table 5.3).

Table 5.3: Station information.

TSA	station name	station code	catchment area [km ²]	observed peak flow [m ³ /s]
Kootenay Lake	Five Mile Creek	08NJ168	47.5	6.6
	Duck Creek	08NH016	57.0	3.4
	Keen Creek	08NH132	92.2	13.4
	Silverton Creek	08NJ005	99.7	12.0
	Lemon Creek	08NJ160	178.0	20.0
	Grohman Creek	08NJ099	88.8	11.0
	Clearwater Creek	08NE116	49.7	5.0
	Akokli Creek	08NH102	50.5	4.0
	Redfish Creek	08NJ061	26.2	3.8
	Carney Creek below Pambrun Creek	08NH131	118.0	17.0
	Meadow Creek above John Creek	08NH124	62.7	6.0
	John Creek near Meadow Creek	08NH125	34.7	4.5
	Glacier Creek near Howser	08NH003	282.0	30.0
Prince George	Kazchek Creek near the mouth	08JE005	881.0	30.0
	Tsilcoh River near the mouth	08JE004	414.0	18.0

The runoff generation model was run for all third order watersheds. The runoff simulations for watersheds larger than third order consisted of the sum of several third order watershed runoff simulations.

5.2.7 Peak Flow Modification

The baseline of our analysis was the situation in 1995, before the Mountain Pine Beetle infestation began. The effect of land use change on peak flow was assessed for two scenarios. Scenario 1 was based on the projected mountain pine beetle infestation for the year 2017 (Eng et al., 2006). It was assumed that the killed trees corresponded to conditions under gray attack, and no salvage logging operation was undertaken. Scenario 2 included the assumptions of scenario 1, except that projected salvage logging on infested areas was incorporated. Salvage logging was assumed in stands with projected tree mortality exceeding 75%, a pine proportion greater 60%, and age older than 80 years. These criteria reflected optimal salvage harvest management conditions (Fall et al., 2003). The effects of the two scenarios on peak flow were assessed by calculating changes in peak flow in each watershed relative to the reference forest condition. The reference condition was derived from the BTM land cover data set (Ministry of Sustainable Resource Management, British Columbia, 2001), which represented the land cover in the year 1995 before the MPB epidemic started.

5.3 Results

5.3.1 Snow Melt Prediction

The results of the snow melt simulation evaluation are presented in Table 5.4. The SWE and consequently snow melt were not as well simulated in the mountainous Kootenay Lake TSA compared to the Prince George TSA. The model overestimated the snow pack on average in the beginning of the melt season in both areas and slightly underestimates SWE in June. The coefficient of determination improved over the melting period in the Kootenay Lake TSA and declined in the Prince George TSA.

Table 5.4: Comparison of observed and simulated mean snow water equivalent (SWE).

			Kootenay Lake TSA			Prince George TSA		
			April	May	June	April	May	June
observed SWE	mean	[mm]	551	543	292	222	168	67
	standard deviation	[mm]	308	382	398	176	209	97
observed vs. simulated SWE	<i>ME</i>	[mm]	-113	-41	13	-56	-26	12
	<i>RMSE</i>	[mm]	244	249	257	87	77	41
	<i>R</i> ²	[-]	0.53	0.67	0.78	0.94	0.87	0.80

5.3.2 Peak Flow Prediction

Observed and predicted mean annual peak flow for the 15 assessed watersheds are shown in Figure 5.4. The simulations slightly underpredicted the observed peak flow. The overall coefficient of determination equaled 0.95. Despite the relatively large range in discharge and drainage area, no clear bias between smaller or larger watersheds was observed.

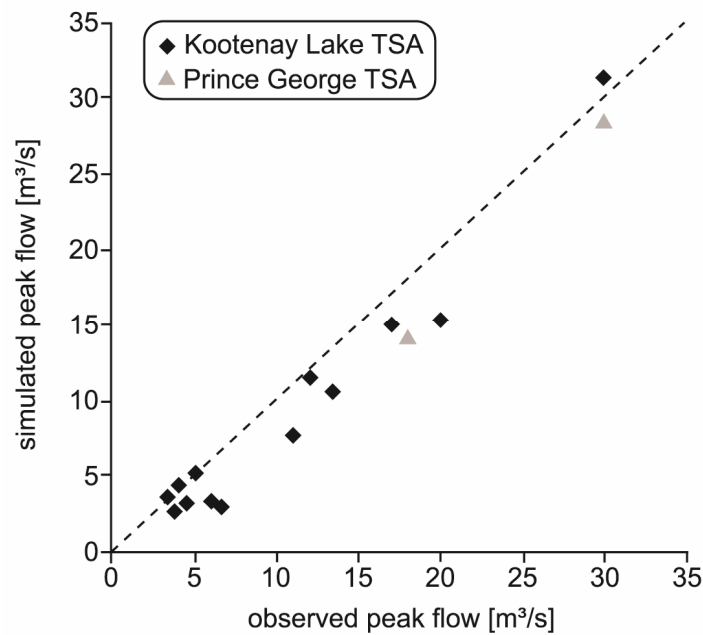


Figure 5.4: Evaluation of the mean annual peak flow predictions. The dashed line represents equal values in observed and simulated peak flow.

5.3.3 Peak Flow Modification

The effects of the landuse changes following the two scenarios varied substantially among watersheds. As an example, the spatial pattern of peak flow changes of third order watersheds of the Prince George TSA is shown in Figure 5.5. The variability in predicted changes across watersheds was considerable. For some watersheds peak flow increases of over 60% were predicted. Other catchments do not exhibit notable flow increases. The patterns of the changes for scenario 2 were clearly related to the planned cut blocks.

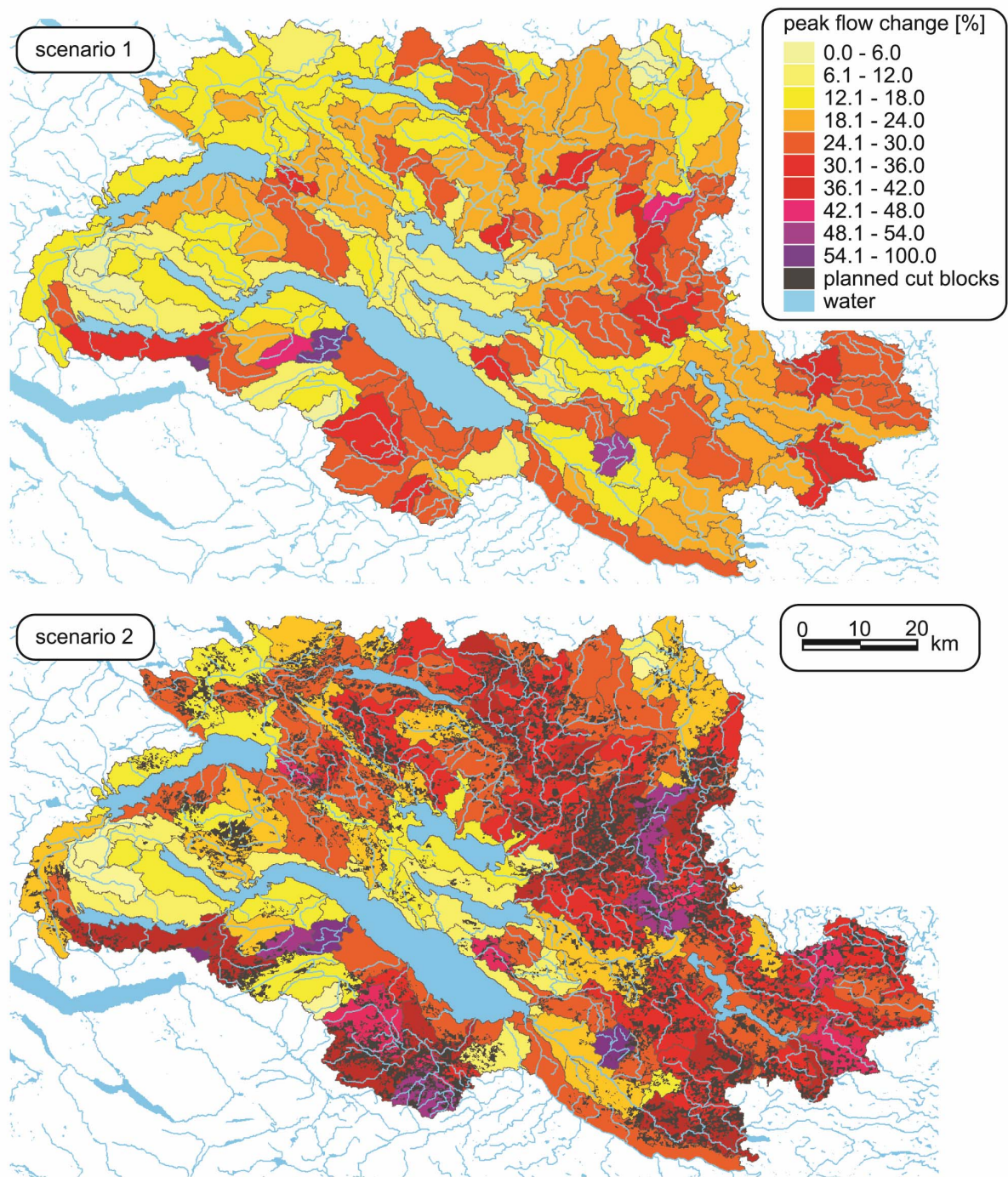


Figure 5.5: Predicted peak flow changes per watershed. Peak flow change predictions according to the two scenarios are shown for the Prince George TSA (longitude of TSA center point: 124° 23' W, latitude: 54° 39' N)

The differences in the variability of peak flow increase between the two study regions and scenarios can be seen in Figure 5.6. For both scenarios, substantially higher peak flow increases were predicted for the Prince George TSA compared to the Kootenay Lake TSA. Due to the

many watersheds with marginal peak flow increases, the distribution for Kootenay Lake was positively skewed in contrast to the fairly symmetric distribution of Prince George. The largest predicted changes were around 20% for scenario 1 and over 60% for scenario 2 for both TSAs.

For up to 78% of the Kootenay Lake TSA, peak flow increases of less than 5% were predicted according to scenario 1. The peak flow distributions of scenario 2 looked similar to those for scenario 1, yet at a much higher level. The peak flow increases of scenario 2 were approximately three times higher than for scenario 1. In the Kootenay Lake TSA the peak flow change of 74% of the watersheds will probably not exceed 20%, whereas in the Prince George TSA approximately 5% of the watersheds will not exceed this level. The average change for all watersheds in the Kootenay Lake TSA for scenario 1 and scenario 2 was 2.9% and 12.3%, respectively; however, the average changes in the Prince George TSA for scenario 1 and scenario 2 were 10.0% and 41.5%, respectively.

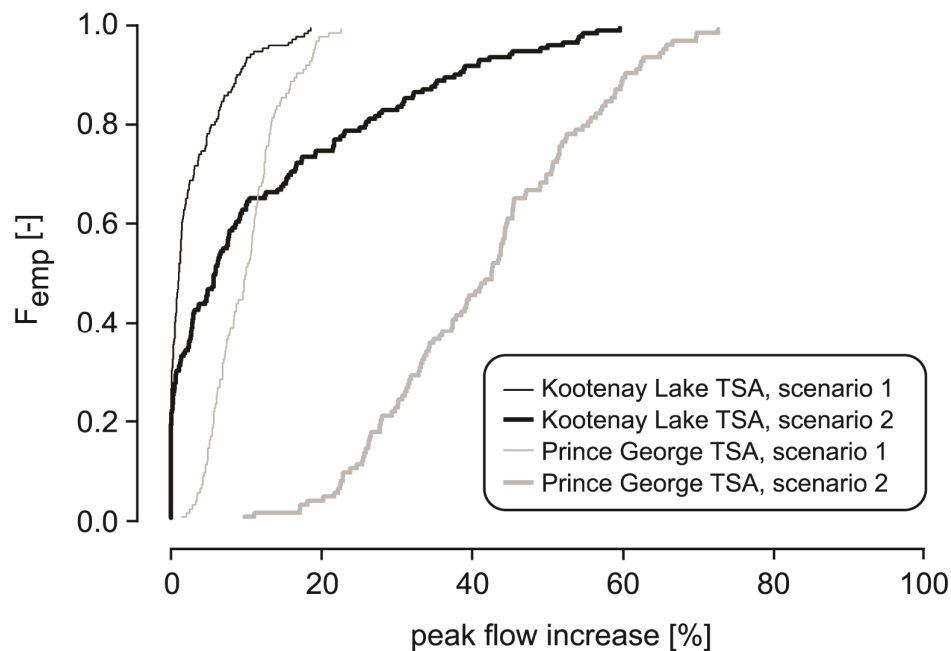


Figure 5.6: Distributions of peak flow increase predictions. Empirical distributions F_{emp} of peak flow increase across watersheds are shown for both study regions and scenarios.

The proposed model directly related changes in vegetation cover to the predicted runoff generation processes and hence to the predicted runoff contribution in each grid cell. This allowed an assessment of the sensitivity of peak flow to land use changes. For each of the 292

watersheds of both TSAs the predicted peak flow changes were related to the area affected by landuse change (Figure 5.7). Land use change areas accounted for the mountain pine beetle infestation (gray attack) or a combination of infestation and forest management (clear cutting).

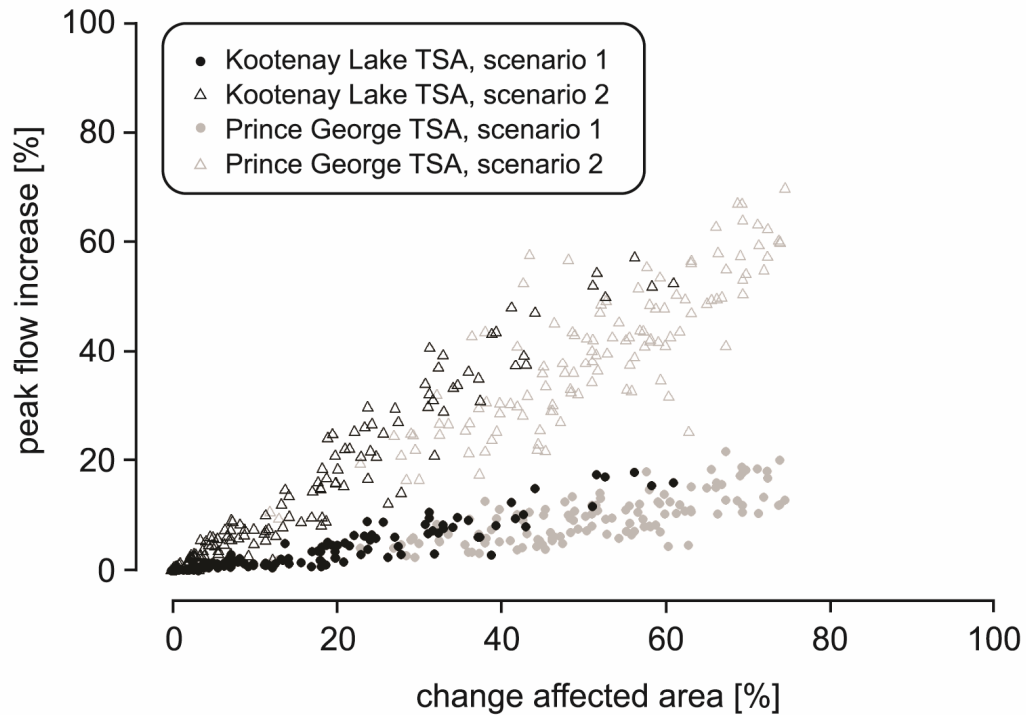


Figure 5.7: Predicted peak flow increase versus landuse change. Predicted peak flow increase is shown in relation to the proportion of area with land-use changes either by gray attack or clear cutting.

For both scenarios a relationship was found between the proportions of change-affected area and predicted peak flow increase. In agreement with scenario definitions, the peak increase per change-affected area was smaller for scenario 1 than 2. However, for both the Kootenay Lake and Prince George TSA similar relations between peak flow increase and landuse change were observed within each scenario. The predicted relationships demonstrated a substantial influence of landuse changes on peak flow increase. For example, land cover change between 40% and 50% was predicted to result in changes in peak flow between 3% and 16% under scenario 1, and between 22.5% and 60% under scenario 2.

5.4 Discussion

5.4.1 Snow Melt Prediction

Over the entire snowmelt period, the proposed model predicted snow water equivalents and subsequent snow melt fluxes reasonably well. Furthermore, the model results compared well with other similar larger-scale, simple snow models in mountainous regions. For example, Kattelmann et al. (1985) modeled SWE with a RMSE ranging from 30 to 610 mm in the Sierra Nevada, USA. The prediction by Kattelmann et al. (1985) was not clearly better than our SWE estimation, even though they used site specific meteorological data and site specific parameterization for snow accumulation and ablation simulations. Our approach with monthly climatic input data is widely accepted in large-scale hydrological models (e.g. Brown et al., 2003). However, the spatial resolution of the chosen climate data sets might be insufficient in mountainous regions of British Columbia with high precipitation variability. The lack of spatial resolution and variability might be a reason for the fairly poor SWE predictions in the Kootenay Lake TSA in April. Furthermore, snow water equivalent of high elevation watersheds in British Columbia could still increase due to snow fall events in April (Jost et al., 2009). Jost et al. (2009) demonstrated a high daily variability of SWE increase and decrease in April. The low temporal resolution of climatic input data, in particular temperature, could be another reason for the poor April snow melt predictions in the Kootenay Lake TSA.

5.4.2 Peak Flow Prediction

The fairly good peak flow simulations (Figure 5.4) indicated that the model structure and parameters for applications in the two TSAs were reasonable. The value of dominant runoff generation process models has been tested with experiments (Scherrer and Naef, 2003) and simulations (Faeh et al., 1997), but mainly in steep terrain. Based on our results, we could speculate that dominant runoff generation processes also had value in the less steep Prince George TSA. Furthermore, we speculated that our approach was successful because peak flow in both TSAs was clearly dominated by snowmelt with minor influences by rain on snow events. However, further research is needed to assess these model types with internal observation or within nested catchment approaches.

5.4.3 Peak Flow Modification

Since most of the watersheds in the Prince George TSA were infested by mountain pine beetle, and large areas were covered by lodgepole pine, substantial peak flow increases were predicted for all watersheds, in particular under scenario 2 (Figure 6). In comparison, many watersheds in the Kootenay Lake TSA were not affected by the infestation since the watersheds were located above tree-line; consequently, no peak flow increase was predicted for more than 20% of this TSA.

Theoretically, the combination of the spatial distribution of landuse change with the spatial distribution of runoff generation areas controlled the peak flow sensitivity to MPB infestation. If one of the factors ‘landuse change’ and ‘runoff generation’ were predominant in a watershed, peak flow sensitivity might be controlled by the other factor. For example, if DRP areas nearly covered entire watersheds with low spatial variability of snowmelt, the level of landuse change might be highly related to the peak flow increase. On the other hand, if landuse were similarly changed across entire watersheds, peak flow sensitivity might be controlled by the spatial distribution of DRP areas and snowmelt characteristics. However, Figure 5.7 suggests primarily the factor ‘landuse change’ and not ‘runoff generation’ controlled peak flow sensitivity. We speculated that this predominance occurred due to the low variability in total DRP area proportion between watersheds in the assessed TSAs. The steep Kootenay Lake TSA exhibited many SSF areas, but few with SOF, HOF, and CHA as the dominant process. In contrast, the flat Prince George TSA had more SOF, but fewer SSF areas than the Prince George TSA. However, the total proportion of DRP areas was similar in both TSA. That could be a reason why landuse changes dominated peak flow increases (Figure 5.7).

We critically evaluated the level of peak flow increase for different change affected areas by comparison with studies reviewed by Moore and Wondzell (2005). Most of the reviewed studies focused on logging treatments rather than insect infestation, and only a few study sites were dominated by lodgepole pine. For example, Troendle and King (1987) found a peak flow increase of 50% for 36% of the watershed clear cut, which clearly exceeded our predicted peak flow increase for a similar affected area and scenario (Figure 5.7, scenario 2) by 10% to 30%. In a different watershed with a clear cut proportion of 10%, Troendle and King (1987) did not observe a significant peak flow increase. This is in agreement with our peak flow increase prediction of less than 10% for a similar changed area. However, Troendle and King (1987)

assessed the impact of landuse change at much higher elevation with different precipitation characteristics, which might substantially influence the effect of vegetation on runoff formation.

Even though treatment, forest type and elevation range of the review by Moore and Wondzell (2005) were not directly comparable to the conditions of our simulations, most stream flow predictions presented by Moore and Wondzell (2005) fell within our prediction range (Figure 5.7). Clearly higher (e.g. Hudson, 2001: 123% peak flow increase for 51% shelter wood cut) or lower predictions (e.g. Cheng et al., 1975: 22% peak flow decrease for 71% clear cut) than our study were rare in Moore and Wondzell (2005). However, Hudson (2001) and Cheng et al. (1975) studied coastal rain-on-snow and rain dominated catchments; that is why their results are not directly comparable to our study. Of higher comparability is a paired catchment study by Moore and Scott (2005) in snowmelt-dominated watersheds in the southern interior of British Columbia. Logging of 27% of the catchment area of 33.9 km² due to a pine beetle infestation resulted in earlier peak flow (0 to 40 days). Peak flows in the first post-harvest decade did substantially increase. On average peak flow increases of approximately 75% according to the pre-harvest paired catchment regression were observed (Moore and Scott, 2006), which clearly exceeds the predictions of our study. However, Moore and Scott (2005) also observed noticeable forest recovery with influence on peak flows; on average over the second post-harvest decade peak flow increases of approximately 8% were observed (Moore and Scott, 2006).

Our results were compared to other simulation studies. Whitaker et al. (2002) found an elevation dependence of peak flow increase due to landuse change. For a scenario with 66% of forest logged at lower elevation with moderate snowmelt influence, a peak flow increase of 12% was predicted for radiation dominated snow melt. For the same proportion of clear cut at high elevation, a peak flow increase of 22% was simulated. Schnorbus et al. (2010) simulated peak flow increases in the Fraser basin in British Columbia; for most watersheds they found smaller peak flow increases than presented in this study.

It should be noted that this study only addressed peak flow increase in third order watersheds. For such catchments with size less than 200 km², it was possible to relate peak flow increases at the outlet of the watershed to change affected area in the watershed. However, routing effects should be considered for peak flow predictions at larger scales. Furthermore, larger catchments would have greater environmental variability and thus more opportunity for desynchronizing

snow melt following forest disturbance. Eaton et al. (2010) reported the effects of a forest fire in the lower parts of a watershed, and inferred substantial snow melt desynchronization. In the post-fire period Eaton et al. (2010) observed more stream flow peaks and clearly earlier and steeper rising limbs of freshet hydrographs. These observations are in correspondence with our model set up, where the peak flow timing due to snow melt depends on spatially distributed snow melt processes (Figure 5.2). Furthermore, our spatially distributed model combines landuse changes with dominant runoff generation processes, since in nature catchment physiography controls if melt synchronization alteration due to landuse change results in peak flow superimposition or counterbalance.

For future research we recommended assessing the proposed model in more detail in regions with high variability in precipitation, temperature, geology, and geomorphology, to quantify controlling effects under varying conditions. Furthermore, continuous input data series should be tested at multiple spatial scales. Our introduced approach has the potential to be extended from snowmelt to rain induced peak flow predictions. Potentially, stream flow predictions due to landuse change could be associated with likelihoods or probability of flow exceedance (Alila et al., 2009; Alila et al. 2010).

5.5 Conclusions

We introduced a framework to predict the effects of spatially explicit landuse changes on peak flow in snow-dominated watersheds. This study found good agreement between the snow melt and peak flow predictions of the proposed method without calibration to field observations. Simulations in two test regions in British Columbia showed a clear relation between that peak flow increase and the proportion of changed area. However, the runoff generation processes produce some variability among watersheds with similar land cover change. At the scale of third order watersheds, the introduced method provides a promising tool for hydrologists and forest management to assess the influence of land cover changes on peak flow. The model can also be used to assess best forest management practices since it is spatially explicit and therefore can implement proposed clear cuts or other management options. The presented method is a reasonable approach to relate detailed stand level knowledge to watershed response predictions.

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6 CONCLUDING CHAPTER

6.1 Chapter Contents and Relationships

6.1.1 Process Area Delineation

In this study, a parsimonious, process-based hydrological model was developed, evaluated, and applied to predict watershed responses with or without landuse changes. The proposed model relied on the widely accepted concept of dominant runoff generation processes (DRP; Scherrer and Naef, 2003). The new model depended on four dominant runoff generation processes, which were channel interception (CHA), Hortonian overland flow (HOF), saturation excess overland flow (SOF), and shallow subsurface flow (SSF).

At the beginning of the DRP model set up, the runoff generation areas of the four DRP had to be delineated with simple model structures. The role of channel interception as a fairly simply simulated process has been well established (e.g. Gomi et al., 2002). In the proposed model, CHA areas were estimated with available stream and lake data. Also, Hortonian overland flow and its dependence on the hydraulic conductivity has been thoroughly examined in previous research studies (e.g. Woolhiser et al., 1996). For example, Ziegler and Giambelluca (1997) demonstrated the generation of Hortonian overland flow on unpaved roads. Since no soil information was for available for the entire province of British Columbia, roads were used as the only criterion to delineate HOF generation areas in this study. In spite of this simplification, the CHA and HOF area delineation of this study did not substantially deviate from classification by Scherrer and Naef (2003) and was probably quite reliable.

In contrast to CHA and HOF, predictions of SOF generation areas involve more controversy. In particular, simple topography-based approximations of saturated areas based on the topographic index (e.g. Guentner et al., 1999) have been subject to divisive discussions (Quinn et al., 1995; Ambroise et al., 1996; Merot et al., 2003). Therefore, saturation predictions of two different topographic indices were tested in chapter 2. Furthermore, a groundwater distance approximation measure was introduced to cope with the variety of saturation processes (riparian or hillslope induced saturation; McGlynn and McDonnell, 2003). Topographic indices and groundwater distances were evaluated separately and in combination. The evaluation of the area delineation

model structures was done at large scale, since climate differences could influence a purely topographic selection of runoff generation areas. Chapter 2 presents test criteria delineations and results of the model evaluation.

Subsurface storm flow was addressed in previous research, in which Whipkey (1965) found slope to be a key control in subsurface storm flow on forested hillslopes. In our approach, slope was incorporated in the DRP area delineation. Other important factors like soil permeability, water-impeding layers, or antecedent wetness could not be estimated for the entire province of British Columbia from topographic data. However, due to the importance of gradient as the driving force of subsurface flow, other hydrologic models have estimated subsurface flow generation from topographic data (e.g. Quinn et al., 1991). The success of such subsurface flow estimation mainly depended on whether the hydraulic gradient could be approximated by the soil surface (e.g. for steep slopes with thin permeable soil layer; Wigmosta and Lettenmaier, 1999). Moreover, topography-based models could not account for macropores, whose relevance for subsurface storm flow in forests is widely accepted (Mosley, 1982). In steep watersheds, this disadvantage might be diminished, since Weiler and Naef (2003) reported a reduced influence of single macropores in subsurface storm flow generation for steep watersheds. Consequently, we assumed in this study that shallow subsurface storm flow was generated at locations with a steep gradient.

The combination of previous research about runoff generation area estimation (CHA, HOF, and SSF) with the saturation prediction evaluation (relevant for SOF, chapter 2) resulted in DRP area delineation criteria for all four processes.

6.1.2 Process Models

Dominant runoff generation process areas have often been delineated (e.g. Schmocker-Fackel et al., 2007). However, only a few investigations transferred DRP area information into a precipitation runoff prediction model (Hellebrand and van den Bos, 2008). The characteristics of dominant runoff generation processes after intensive precipitation have been described in detail (Scherrer et al., 2007). But the process knowledge has rarely been used for parsimonious process-based, dynamic models. In particular, connectivity between dominant runoff generation process areas has not previously been studied.

Consequently, simple DRP models and connectivity structures were introduced and evaluated in chapter 3. The evaluation of entire model structures provided new insights for each of the model subcomponents (from DRP areas and process models to process interactions). The result of chapter 3 was the development of a comprehensive DRP framework which could be used for predictions (chapters 4 and 5).

6.1.3 Predictions

Hortonian overland flow and saturation excess overland flow have been expected to rapidly contribute to storm runoff after intensive or long-lasting storms (Naef et al., 2002). However, delayed flow reaction has been observed in watersheds with high subsurface flow area proportions (Scherrer, 1997). The idea behind chapter 4 was that empirically observed DRP effects on runoff dynamics could possibly be used for stream flow predictions. In a first step of chapter 4, dominant runoff generation areas were developed according the concepts and thresholds gained in chapters 2 and 3. Second, the runoff prediction value of DRP models was evaluated given additional topography, landuse or climate data.

In chapter 5, the DRP concept was tested for its ability to predict the effects of landuse changes on peak flow. It has been demonstrated that sensitivity of streamflow to landuse changes depends on how much they affect dominant runoff generation processes (Naef et al., 2002; Bronstert et al., 2002). The first step of chapter 5 consisted of the evaluation of DRP-based peak flow predictions. In a second step, the effects of landuse change scenarios on peak flow were predicted with a dominant runoff generation process framework.

6.2 Study Findings

Table 6.1 presents the most important findings of this study. Due to the substantial topic variability among chapters, the findings are grouped by chapter. Furthermore, this study generated findings and procedures that have not been discussed in previous research, but were not directly related to the objectives of this study; the most relevant side effect findings are shown in Table 6.2.

Table 6.1: Major study findings.

Chapter	Findings
2	<ul style="list-style-type: none"> • Saturation footprint indices agreed fairly well with observed saturation. • Across climates, saturation foot print indices need to be differentiated into primary (minor climate dependence; e.g. plants. hydromorphic features) and secondary footprint indices (major climate dependence; e.g.: morphology and soil information). • The best agreement between observed and predicted saturation was achieved for a model structure which consisted of a combination of topographic index and vertical distance to the groundwater. • Climate dependence of optimized model parameters was found for three out of four assessed watersheds.
3	<ul style="list-style-type: none"> • DRP areas were redistributed toward locations along flow paths if connectivity was considered. • Based on model fit and parameter feasibility it was more meaningful to incorporate connectivity in process models than in simple runoff coefficient models. • Process models generated more realistic temporal connectivity distributions than runoff coefficient models. • Hortonian overland flow areas were more affected by connectivity than saturation excess overland flow or subsurface storm flow areas.
4	<ul style="list-style-type: none"> • DRP area distributions per se explained substantial amounts of the variability of snowmelt dominated catchment response measures (adjusted coefficient of determination greater than 0.8). • In addition to topographic data. DRP were most useful for the prediction of rain storm dictated response measures. • The additional utility of climate data was surprisingly low, particularly for snowmelt dominated measures.
5	<ul style="list-style-type: none"> • Snowmelt and peak flow predictions based on DRP agreed well with observed data. • Predicted peak flow increase was related to the landuse change-affected area proportion. • Peak flow changes up to 70% were predicted. • Predicted peak flow increases were approximately three times higher for the scenario with salvage logging compared to the scenario based on Mountain Pine Beetle infestation without logging.

Table 6.2: Most important side effects of this study.

Chapter	Findings
2	<ul style="list-style-type: none"> • Not Kappa statistics (which have often been applied) but simple agreement measures should be used to evaluate spatially distributed saturation predictions with observed saturation patterns. • Topographic index calculation: we suggest adding two degrees to the slope to cope with flat areas. • The saturated proportion of the observed watershed increased with storm intensity. However, at most in 20% of the watershed saturation was found to be saturated even for very intensive storms. These findings advocated the practicability of the variable source area concept in the study area, which was a fundamental prerequisite for DRP approaches. • Microtopography was estimated with a roughness height measure, which was combined with concavity and slope to assess morphology induced soil saturation.
3	<ul style="list-style-type: none"> • A framework was developed how to estimate parameters and evaluate performances of comprehensive DRP models (areas, processes, connectivity).
4	<ul style="list-style-type: none"> • High correlations of were found between daily flow increase and decrease across watersheds. • A seamless, hydrological meaningful 25m digital elevation model was generated for the entire province of British Columbia.
5	<ul style="list-style-type: none"> • A concept was delineated how to predict peak flow volume and timing in snowmelt dominated watersheds.

6.3 Limitations and Future Research

It should be noted that the field work part of this study concentrated on soil saturation and did not include saturation overland flow, since the former could simply be assessed with saturation samplers. It was assumed that soil saturation areas could be used to estimate saturation overland flow generation areas. The field evaluation was restricted to SOF; CHA, HOF, and SSF model structures were defined based on previous research. Due to a lack of soil data, there was no response characteristic based division of each process (e.g. HOF1 and HOF2; Scherrer and Naef, 2003). As a result, the processes of this study represented average response properties of each DRP. The routing procedures of the introduced DRP model structures were restricted to hillslope water movement; however, routing in the channel was neglected under the assumption of much shorter travel time in the stream compared to the hillslope. The connectivity concept was limited to connection of contributing areas and the channel network; cell-to-cell water transfer algorithms - which particularly included no-DRP cells - were not assessed in this study in order to comply with the parsimonious model structures.

Peak flow change prediction due to landuse alteration concentrated on snowmelt dominated watersheds; catchments with rainfall induced peak flow were not assessed. Furthermore, only 25 m digital elevation models were assessed; due to the lack of high resolution digital elevation

models for entire British Columbia higher resolution DEMs could not be considered, even though optimum terrain-based hydrologic simulations have been achieved for resolutions of approximately 10 m (Jencso et al., 2009) or higher (Thompson and Moore, 1996). However, the DEM resolution should not be too high either to prevent overestimation of small scale morphological patterns (Lane et al., 2004).

In future studies, the introduced model as well as its subcomponents should be evaluated with respect to their sensitivity toward precipitation, landuse, soil type, and geology. These factors should vary substantially among the selected watersheds, which allow defining conditions within which DRP models provided reasonable predictions, but also applicability limits of DRP models (e.g. DRP models should not be applied to karst formations). In particular, predicted source areas should be evaluated with isotope studies. Continuous time series from different source areas should be used to evaluate DRP predictions. Furthermore, we suggest using runoff data in British Columbia before, during, and after the mountain pine beetle infestation to evaluate DRP predictions of the MPB infestation on peak flow.

6.4 Significance to the Research Field

This study introduced a framework for developing, applying, and evaluating a hydrologic model based on the concept of dominant runoff generation processes in hydrological modeling. The models were used to predict catchment response and landuse change effects on peak flow sensitivity. A framework was introduced in this study how to predict the spatial extent and dynamic response characteristics of dominant runoff generation process areas. The findings implied that process interactions and connectivity to the channel network should be considered in DRP models. The study confirms the usefulness of DRP concepts in predicting catchment response and landuse change effects.

Dominant runoff generation process concepts can be applied in peak flow hazard models. They provide process-based decision support for flood prediction management with particular potential in ungauged basins. DRP models could be used to incorporate stream flow sensitivity to landuse change in forest clear cut planning. For the Fraser River basin (230,000 km²) in British Columbia, a DRP-based peak-flow hazard model has been developed (Carver et al., 2009). The model was to be used in risk-based hydrologic modeling to produce a comprehensive knowledge of mountain pine beetle-infestation effects on the hydrology of the Fraser River basin and its

major sub-basins. Distinct allocations of source areas contributing to storm runoff by DRP concepts have a high potential in ecological projects. For example, logging-related sediment generation, which impeded endangered fish species reproduction in a small watershed in British Columbia, has been estimated with DRP models (Wood et al., 2008). Moreover, DRP concepts have started to be used in water quality research (Frey et al., 2009; Brown and Schreier, 2009), and cross-disciplinary studies with social sciences (Kim et al., 2008; Kim et al., 2010). These applications indicate the potential of dominant runoff generation process concepts for comprehensive hydrological and ecological predictions of landuse change effects. In contrast to traditional hydrologic models, the proposed model framework incorporates change prediction characteristics:

- **Calibration and evaluation:** Final DRP models are neither calibrated nor validated with data. The model development is based on process knowledge.
- **Dominating processes:** Characteristics of dominant runoff generation processes such as Hortonian overland flow, saturation excess overland flow, and subsurface flow are extracted from field studies.
- **Scales:** Processes are identified at hillslope scale to make predictions at basin or regional scale. The programming code structure of the model allows predictions for large basins (e.g. the Fraser basin; 248,000 km²) or the entire Province of British Columbia (945,000 km²).
- **Threshold and cumulative effects:** The model set up considers threshold process behavior (e.g. for infiltration or soil saturation). Cumulative effects (e.g. snow melt rate, timing, and precipitation on snow) are considered.
- **Spatially-distributed patterns:** The model is entirely spatially distributed.
- **Landuse-hydrology interaction:** Hydrological processes are coupled with landuse to allow change predictions.
- **Data sources:** The use of easily available elevation data sources has been evaluated, and compared to the benefit of other data sets (e.g. climate data) for runoff predictions.
- **Cross-disciplinarity:** The model set-up emerged from knowledge and experience from soil science, forestry, snow and hillslope hydrology. The predictions have been applied to water quality and social studies.

To sum up, this study presents a new, parsimonious framework how future hydrologic models could make spatially-distributed predictions of landuse change effects without calibration with historical data.

6.5 References

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