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Application of TopoFlow, a spatially distributed hydrological model, to the Imnavait Creek watershed, Alaska

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[1] This study presents the application of the hydrological model TopoFlow to the Imnavait Creek watershed, Alaska, United States. It summarizes the hydrologically important processes in this arctic basin, and focuses on the modeling of the hydrological processes in 2001. The model is evaluated for its capability to reproduce the different components of the hydrological cycle. Model simulations are done for different climate change scenarios to evaluate the impacts on the hydrology. Imnavait Creek ($\sim$2 km^2) is underlain by continuous permafrost, and two features characterize the channel network: The stream is beaded, and numerous water tracks are distributed along the hillslopes. These facts, together with the constraint of the subsurface system to the shallow active layer, strongly influence the runoff response to rain or snowmelt. Climatic conditions vary greatly during the course of the year, providing a good testing of model capabilities. Simulation results indicate that the model performs quantitatively well. The different components of the water cycle are represented in the model, with refinements possible in the small-scale, short-term reproduction of storage-related processes, such as the beaded stream system, the spatial variability of the active layer depth, and the complex soil moisture distribution. The simulation of snow melt discharge could be improved by incorporating an algorithm for the snow-damming process.

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1. Introduction

1.1. Hydrology of the Arctic

[2] The presence of permafrost is the primary factor distinguishing arctic from temperate watersheds. Permafrost underlies approximately 24% of the exposed land area in the Northern Hemisphere, making it a significant proportion of the land mass [*Romanovsky et al.*, 2002]. The permafrost condition is a crucial component in its influence on many of the hydrologic processes in the arctic and subarctic environments. The presence of permafrost significantly alters surface and subsurface water fluxes, as well as vegetative functions [*Walsh et al.*, 2005]. Permafrost dominates microclimatology and the thermal regime, including evapotranspiration [*Hinzman et al.*, 1996, 2006]. Permafrost controls water storage processes and the energy and water balances [*Boike et al.*, 1998; *Bowling et al.*, 2003].

[3] *Hinzman et al.* [2005] point out that the primary control on hydrological processes is dictated by the presence or absence of permafrost, but is also influenced by the thickness of the active layer, the thin layer of soil overlying permafrost that thaws in the summer. The active layer in the arctic varies from several tens of centimeters to 1 or 2 m in depth. It is of pivotal importance, as most hydrological and biogeochemical processes occur in this zone [*Kane et al.*, 1991a; *Walsh et al.*, 2005]. The conditions for plant growth, gas fluxes, groundwater flow regimes, and soil formation are all limited and to some extent determined by the active layer [*Boike et al.*, 1998]. The permafrost beneath the active layer limits the amount of soil water percolation and subsurface storage of water [*Vörösmarty et al.*, 2001]. Whereas nonpermafrost soils allow a deep groundwater system, the subsurface movement of water in permafrost-affected soils is largely confined to the shallow active layer. Therefore lateral flow is more important than in nonpermafrost soils [*Slaughter and Kane*, 1979]. These characteristics have a large impact on the runoff response. Permafrost generally accelerates the initiation of runoff [*McNamara et al.*, 1998]. As the water movement through the near-surface soils is relatively fast, the runoff response to precipitation is characterized by a rapid rise to peak flow and a rapid decline following peak flow [*Dingman*, 1973]. In addition, response times are shortened because vegetation in these areas tends to be sparse [*Church*, 1974]. While permafrost-dominated watersheds generally have a larger contributing area and a higher specific discharge, the specific base flow is lower compared to nonpermafrost regions [*McNamara et al.*, 1998].

[4] The annual thawing and freezing of the active layer are the driving forces for many surficial processes, such as

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cryoturbation. These perennial processes also have a control on the hydraulic properties of the soil, specifically the storage capacity and hydraulic conductivity [*Hinzman et al.*, 1991]. The variation of hydraulic properties results in runoff patterns which change throughout the thaw season. To understand hydrologic dynamics of the arctic, it is conducive to study the seasonal change in soil moisture in the active layer. An overview over the seasonal active layer characteristics is given in section 2.

1.2. Arctic Hydrology in a Changing Climate

[5] Air temperature, snow cover, and vegetation, all of which are affected by climate change, affect the temperature of the frozen ground and the depth of seasonal thawing. In interior Alaska, United States, the warmer climate has led to shrinking permafrost coverage and an increased active layer depth [*Osterkamp and Romanovsky*, 1999].

[6] General circulation models predict that the effects of anthropogenic greenhouse warming will be amplified in the northern high latitudes due to feedbacks in which variations in snow and sea ice extent, the stability of the lower troposphere, and thawing of permafrost play key roles [*Serreze et al.*, 2000]. Over the next 100 a the observed changes are projected to continue and their rate to increase, with permafrost degradation estimated to occur over 10–20% of the present permafrost area, and the southern limit of permafrost expected to shift northward by several hundred kilometers [*ACIA*, 2004].

[7] A progressive increase in the depth of seasonal thawing could be a relatively short-term reaction to climate change in permafrost regions, since it does not involve any lags associated with the thermal inertia of the climate/permafrost system [*Walsh et al.*, 2005]. There is a general consensus among models that seasonal thaw depths are likely to increase by more than 50% in the northernmost permafrost locations [*Walsh et al.*, 2005]. It appears that first-order impacts to the arctic, expected with a warming climate, result from a longer thawing/summer period combined with increased precipitation [*McCarthy et al.*, 2001]. The longer snow-free season and greater winter insulation produce secondary impacts that could cause deeper thaw of the active layer or greater melt of permanently frozen ice in glaciers and permafrost, increased biological activity, and changes in vegetative communities. Tertiary impacts arise as animals, people, and industry respond to the changing ecosystem.

[8] It is crucial to study the impacts of a changing climate on arctic water balances, as many processes are directly or indirectly influenced by components of the hydrological cycle, e.g., soil moisture, runoff, and evapotranspiration. However, the question if the arctic tundra will get wetter or drier is not a simple one as all the components interact with each other. In the Siberian arctic, for example, there is evidence of decreasing lake abundance despite increases in precipitation [*Smith et al.*, 2005].

[9] Changes to the water balance of northern wetlands are especially important because most wetlands in permafrost regions are peatlands, which may absorb or emit carbon depending on the depth of the water table [*ACIA*, 2004; *Walsh et al.*, 2005]. In this way, hydrologic changes will have global implications. Other important feedbacks to global warming are the albedo feedback and the weakening of the thermohaline circulation caused by increased freshwater flux into the Arctic Ocean.

1.3. Objective

[10] The study presented here is an application of the TopoFlow model, described in detail by *Bolton* [2006]. Our objective is to evaluate its capability of representing arctic hydrological processes. First, the hydrologically important processes of Imnavait Creek are described. The study then focuses on comparing the physical hydrology, measured and observed in the field, with model results. The model is executed and evaluated for its capability to reproduce the different components of the hydrological cycle.

2. Site Description

[11] The Imnavait Creek watershed is a small headwater basin of approximately 2 km^2, located in the northern foothills of the Brooks Range (68°30′N, 149°15′W), 250 km south of the Arctic Ocean (Figure 1). The Imnavait Creek flows parallel to the Kuparuk River for 12 km before it joins the Kuparuk River that drains into the Arctic Ocean. The elevation in this area ranges from 880 m at the outlet to 960 m at the southern headwaters. The area is underlain by continuous permafrost, and the topography consists of low rolling piedmont hills. Imnavait Creek has been intensively studied since 1985 by the Water and Environmental Research Center (WERC) at the University of Alaska, Fairbanks. This research is documented in, e.g., *Hinzman et al.* [1991, 1996], *Walker et al.* [1989], *Kane et al.* [1989, 1990, 1991b], and *McNamara* [1997].

[12] If not otherwise specified, all data reported in this section are documented by *Hinzman et al.* [1996]. In the Imnavait Creek watershed the mean annual temperature averages −7.4°C. In January (July) the average air temperature is −17°C (9.4°C). The annual precipitation averages 340 mm, two-thirds of which falls during the summer months of June, July, and August. Most rainfall is light (82% <1 mm/h) and appears evenly distributed over the catchment. Because of the influence of wind and topography, snow distribution and snow pack volumes in the Imnavait watershed are extremely variable both in time (year to year) and space (within the watershed), ranging from a few centimeters on windswept ridgetops to more than 1 m in the valley bottom. Winter snow accumulation generally starts around mid-September. A 20-a record shows that the annual snow water equivalent (SWE) in Imnavait Creek varies from 69 to 185 mm [*Berezovskaya et al.*, 2005]. Snowmelt is initiated between 1 and 27 May and is completed within 6–22 d. This reveals a considerable range in timing of snowmelt initiation. The vegetation is mostly water-tolerant plants such as tussock sedges and mosses [*Walker et al.*, 1989]. Generally, with a relatively impervious barrier so close to the surface, wet conditions exist in the active layer near the surface. This provides the conditions suitable for substantial evapotranspiration during the summer thawing months [*Kane et al.*, 1989].

[13] Imnavait Creek is a north draining, first-order stream. The stream is beaded, meaning that the channel connects numerous interspersed small ponds. These ponds are on the order of 2 m deep and a few meters in length and width [*Kane et al.*, 2000]; see Figure 1.

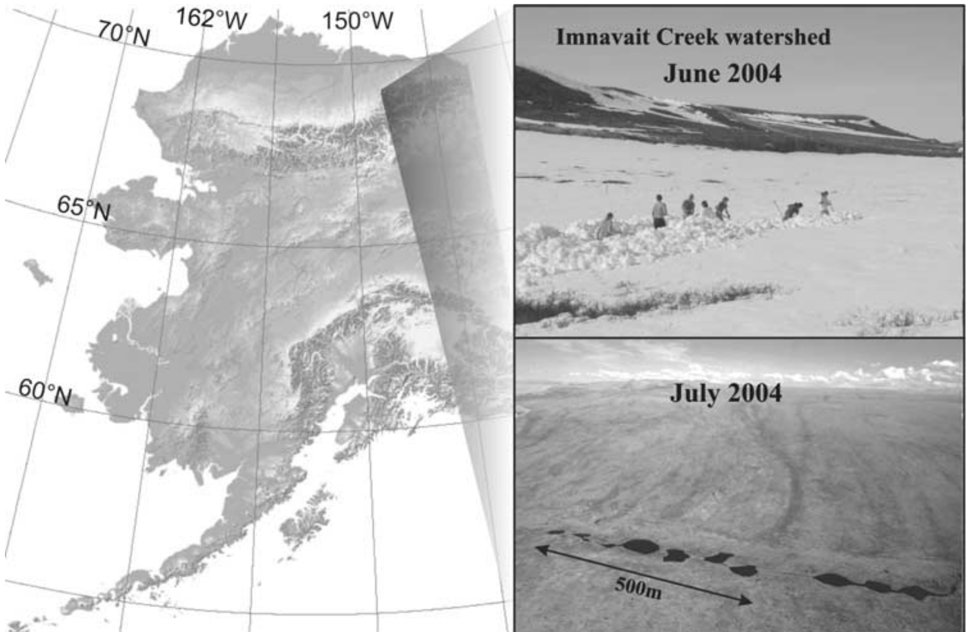

Figure 1. Map of Alaska, United States, with the location of the study area Imnavait Creek.

[14] The headwaters of the creek are found in a nearly level string bog, or strangmoor, with many poorly defined and interconnecting waterways [*Oswood et al.*, 1989]. Along the hillslopes, small drainage channels, or water tracks, carry water off of the slopes down to the valley bottom. The water tracks can be described as shrubby corridors with a width of ∼2 m and spaced at ∼10–20 m along the hillslope. The water tracks contain a system of interconnected deepenings, or small channels of ∼5–10 cm width, which are partly directed parallel to the hillslope. Here the water flow follows microtopographic features, such as tussocks and hummocks [*Quinton et al.*, 2000]. Although quite obvious in aerial photographs, most of these water tracks are difficult to detect on the ground, as they are not incised [*Hastings et al.*, 1989; *McNamara*, 1997]. The water tracks generally take the most direct route down the slope but do not connect directly with the stream in the valley bottom. As the slope flattens out in the valley bottom, water moving down the water tracks disperses into numerous poorly defined channels and slowly makes its way over to the creek. Water moves downslope in these water tracks more rapidly than by subsurface means [*Kane et al.*, 1989].

[15] Runoff leaving the basin is usually confined to a period of 4 months, beginning during the snowmelt period in late May until freeze-up in September. Spring runoff is usually the dominant hydrological event of the year [*Kane and Hinzman*, 1988], typically producing the annual peak flow, and about 50% of the total annual runoff volume. Stream flow almost ceases after extended periods of low precipitation, whereas intense summer rainfall events produce substantial stream flow. Whether runoff is produced from rainfall events during the summer is strongly related to rain intensity and duration and antecedent soil moisture conditions [*Kane et al.*, 1989]. Furthermore, the runoff response depends on the snow cover (see section 5.2), the state of the active layer, and mechanisms related to the channel network: In a beaded stream system, small ponds act as reservoirs and can store water intermediately. This mechanism will, depending on the water level of each pond, result in a delayed hydrograph signal. Furthermore, the state of the active layer plays a pivotal role in altering runoff response. The maximum depth of thaw ranges from 25 to 100 cm, severely limiting the ability of the active layer to store large quantities of groundwater. The rate of thaw is dependent upon a number of factors, such as soil properties, soil moisture and ice content, and the distribution and duration of the snow cover. As a result, the depth of the active layer and thus the soil moisture is highly variable both in space and time [*Woo and Steer*, 1983; *Woo*, 1986]. Because of the excessive water supply from snowmelt, the water table in flatter areas rises above the ground surface to generate surface flow. Spring is therefore the time when the extent of surface flow is typically at a maximum. As summer progresses, the soil moisture content is reduced by an increasing depth of thaw and a continued evapotranspiration. This leads to a rapid depletion of the overall soil moisture content, and a nonsaturated zone develops. Occasional heavy rainstorms, however, can revive surface flow [*Woo and Steer*, 1983], and late summer and early fall rainstorms provide a recharge of soil moisture.

3. Models

3.1. Previous Studies

[16] At present, climate models do not represent the soil layers at high enough resolution to achieve the soil output needed to assess changes in permafrost distribution and active layer characteristics. The need for additional detail is particularly great for areas with thin or discontinuous permafrost [*Walsh et al.*, 2005]. Furthermore, the majority of land surface models have been primarily designed for lower latitudes and as such are not capable of realistically simulating the physical processes operating in the extreme

climate of the arctic. However, increasing efforts have been made to adequately model arctic environments over the last 2 decades. Several modeling studies with varying focuses have been applied to the Imnavait Creek watershed, where field data from multiple-year studies are available.

[17] *Hinzman and Kane* [1992] studied the potential hydrological response during a period of global warming using the HBV model. The original version of this model was developed in 1975 by the Swedish Meteorological and Hydrological Institute as a conceptual runoff model and modified for cold regions use by *Bergström* [1976]. It can be described as a reservoir-type model with routines for snowmelt, soil moisture accounting, control of surface and subsurface hillslope runoff response, and a transformation function to handle stream routing. The model input data are observations of air temperature, precipitation, and estimates of evapotranspiration. Model outputs are snowmelt runoff and the entire summer runoff response. Despite of the good congruence of measured and simulated hydrographs the authors report several shortcomings: First, the lack of physically based routines queries its capability of evaluating future changes. Second, the prediction capability could be improved by incorporating the redistribution of snow by winds and the retardation of runoff by snow damming [*Hinzman and Kane*, 1992].

[18] Another model was applied to the same study area by *Stieglitz et al.* [1999]. The simple land surface model TOPMODEL was used to explore the dynamics of the hydrologic cycle operating in arctic tundra regions. The model accounts for the topographic control of surface hydrology, ground thermal processes, and snow physics. This approach relies only on the statistics of the topography rather than its details. This has the advantage of being computationally inexpensive and compatible with the large spatial scales of today's climate models. However, the authors report several deficiencies, such as that the model performance in temperate watersheds is superior to that for arctic watersheds. This is attributed to the neglect of snow heterogeneity, which poses a real obstacle toward application on an arctic-wide basis. Furthermore, the representation of a seasonally changing connectivity of waterways (e.g. the beaded stream system) is seen to be difficult on a statistical base. As such, TOPMODEL is capable of simulating the overall balances, but shortcomings exist in the hydrograph simulation and soil moisture heterogeneity with high temporal resolution.

[19] A third modeling study with an application to Imnavait Creek is presented by *Zhang et al.* [2000]. Here a process-based, spatially distributed hydrological model is developed to quantitatively simulate the energy and mass transfer processes and their interactions within arctic regions (Arctic Hydrological and Thermal Model (ARHYTHM)). The model is the first of this kind for areas of continuous permafrost and consists of two parts: the delineation of the watershed drainage network and the simulation of hydrological processes. The last include energy-related processes such as snowmelt, ground thawing, and evapotranspiration. The model simulates the dynamic interactions of each of these processes and can predict spatially distributed snowmelt, soil moisture, and evapotranspiration over a watershed as well as discharge in any specified channels. Results from the application of this model demonstrate that spatially distributed models have the potential for improving our understanding of hydrology for certain settings. Nevertheless, the authors point out that an algorithm for snow damming, the usage of a higher resolution, and a better data collection network could improve the model results. Furthermore, the use of triangular elements makes it difficult to compare simulation results with other (e.g., remotely sensed) data sets.

[20] From former studies it becomes evident that topography plays a crucial role in the development of soil moisture heterogeneity. The fact that the impacts of this heterogeneity on surface water and energy fluxes are critical and perhaps overwhelming [*Stieglitz et al.*, 1999] leads to the conclusion that the representation of topographic features in a model cannot be neglected. Furthermore, there exist problems in the current models to handle the rapidly changing thermal (permafrost versus nonpermafrost and active layer development) and hydraulic (hydraulic conductivity and storage capacity) conditions typical of the (sub)arctic regime [*Bolton et al.*, 2000].

3.2. TopoFlow

[21] TopoFlow is a spatially distributed, process-based hydrological model, primarily designed for arctic and subarctic watersheds. TopoFlow is primarily based upon the merger of the ARHYTHM model [*Hinzman et al.*, 1995] and a D8-based rainfall-runoff model. Structurally, the most significant differences between the ARHYTHM and TopoFlow models are the incorporation of rectangular elements and flow routing using the D8 method. In the D8 method, horizontal water fluxes occur from one element to one of the eight adjacent elements in the direction of the steepest slope [*O'Callaghan and Mark*, 1984]. The model domain is defined by a rectangular, regular network DEM that encompasses the catchment area. Each TopoFlow element has dimensions of the DEM pixel (x and y directions) with up to ten user-specified layers of variable thickness in the z direction. On the basis of the conservation of mass principal, TopoFlow simulates major processes of the water balance (precipitation, snowmelt, evapotranspiration, groundwater flow, and overland/channel flow) as well as some storage processes (snow accumulation and infiltration/percolation). Most of these hydrologic processes are formulated in the exact manner as the ARHYTHM model and are well documented by *Zhang et al.* [2000]. Yet important improvements have been made in the process simulation component of the model. These improvements include expansion of the methods available to simulate the infiltration and channel flow processes, the ability to handle a variety of input variable formats, and a user-friendly interface. A detailed description of the model structure and the additional methods incorporated into TopoFlow can be found in the work of *Bolton* [2006].

[22] The development of soil moisture heterogeneity and its correct reproduction in models is crucial for the evaluation of its impacts on surface water and energy fluxes [*Boike et al.*, 1998]. TopoFlow addresses these issues through (1) its spatial distribution that explicitly models the movement of water from element to element; (2) by the implementation of physical routines that are unique in cold regions; (3) by providing user-friendly preprocessing tools that aid in handling the spatial variability, such as the

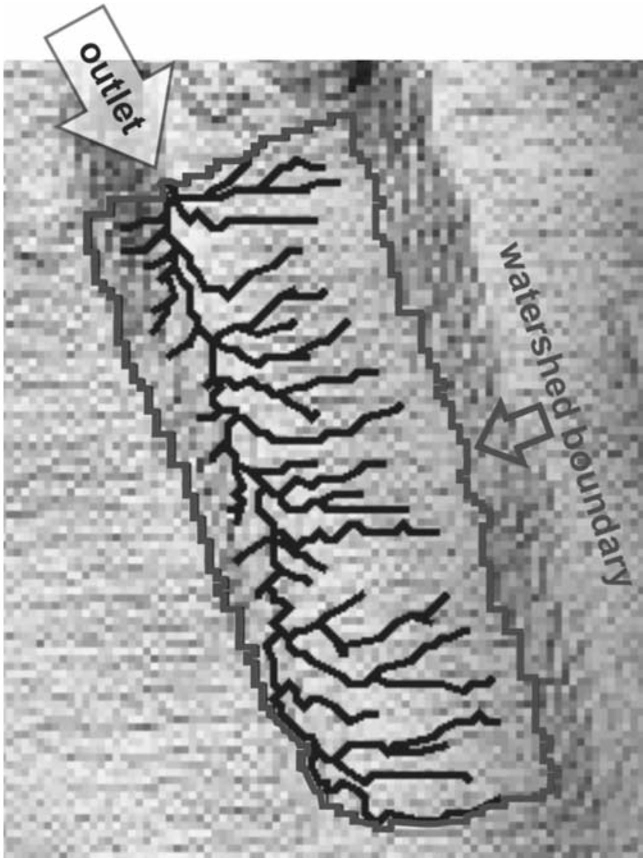

Figure 2. Digital Elevation Model of the Imnavait Creek watershed, its channel network, and watershed boundaries.

distribution of permafrost versus nonpermafrost, the active layer depth, and the snow pack distribution; and (4) by providing a flexible structure that allows the user deal with different data types or the lack of measured parameters.

[23] The hydrological simulation is initiated some hours prior to snowmelt with the end of winter snow pack distribution used as input. TopoFlow supports the degree day and the energy balance method for snowmelt. For evapotranspiration, two methods are provided to account for different availability of input data: the physically based energy balance and the semiempirical Priestley-Taylor approach. TopoFlow allows the spatial distribution of important parameters, such as meteorological variables or coefficients, soil moisture content, soil parameters, and snow pack distribution.

[24] At the time of this study an instantaneous infiltration method was available, and the three different flow processes (channel flow, overland flow, and subsurface flow in the shallow active layer) were incorporated into the model with Darcy's law and Manning's equation [*Schramm*, 2005]. Further improvements of the infiltration and percolation process, such as the finite difference solution of the Richards equation, Green-Ampt, and Smith-Parlange, have recently been incorporated [*Bolton*, 2006] (TopoFlow Web site, http://instaar.colorado.edu/topoflow/).

[25] The active layer starts thawing after snowmelt, continues to thaw during the summer, and reaches its maximum thickness in autumn. Therefore the soil depth in Darcy's equation potentially changes with each time step. Soil moisture capacities for each soil layer also change, because they are related to the soil depth. As the hydraulic conductivity is different for the frozen and the unfrozen soil, flow rates in the frozen layers differ significantly from those in the unfrozen soil. The thawing of the active layer is currently incorporated by a simple square root of time function [*Hinzman et al.*, 1990].

[26] For the overland and channel flow, Manning's formula is used, where the roughness parameter, the shape of the cross section, and the channel width can be specified by the user for each stream order.

4. Model Application

4.1. Digital Elevation Model (DEM)

[27] A DEM with a pixel size of 25 $\times$ 25 m is used in this study. In order to create the input files necessary for TopoFlow simulations, the hydrological software package RiverTools is used in this study. RiverTools defines computationally the watershed area that contributes to a user-specified element. In this study a watershed area of 1.9 km^2 was calculated. This is in good agreement with the manual delineation of 2.2 km^2, taking into account that the headwaters are complex topographically, i.e., a very flat area, and therefore the southern watershed boundary is difficult to determine visually and/or by way of calculation. Figure 2 depicts the DEM of the Imnavait Creek watershed, its channel network, and watershed boundaries.

[28] The DEM is used in RiverTools to generate several files that are needed to extract information for a river network. The flow grid indicates the direction in which water would flow away from the corresponding pixel in the DEM. Here RiverTools provides special algorithms to determine the flow direction in flat areas that are common in the arctic tundra. Furthermore, a RiverTools treefile is derived from the flow grid. This vector-formatted file stores data for the basin such as contributing area and relief. These attributes are stored for every element in a given basin.

[29] In order to differentiate where channel flow and overland flow processes occur, the simulated channel network is compared to the physical system. Elements with a stream order of less than 3 are considered to be overland flow, and those $\geq$3 are locations where channel flow is present. Considering the water tracks (described in section 2) to be channels, the simulated river network compares well with the channel structure that is visible in aerial pictures. Finally, grids of upstream areas, downstream slopes, and Horton-Strahler order are produced with RiverTools for further use with TopoFlow.

4.2. Input Data

[30] Various research projects on the North Slope of Alaska have, since the mid 1980s, resulted in the establishment of several unmanned meteorological and research sites on a north–south transect located along the Dalton Highway. The measurement program is maintained by WERC, and data are available on the WERC Web site (http://www.uaf.edu/water). In the Imnavait Creek basin there are four main sites where data collection takes place: Imnavait basin (68°36′N, 149°18′W, 937 m); Imnavait ridge (68°37′N, 149°19′W, 880 m); Imnavait valley (68°37′N,

Table 1. Soil Parameters of Imnavait Creek Used as Model Input[a]

Soil Layer Depth, cm	Porosity, %	Hydraulic Conductivity, 10^4 m/s
0–10	0.88	1.50
10–20	0.63	0.35
20–30	0.50	0.35
30–40	0.48	0.10
40–permafrost table	0.40	0.10

[a]Data based on *Hinzman et al.* [1991].

149°19′W, 876 m); and Imnavait flume station (68°37′N, 149°19′W, 881 m). Compared with other arctic research basins an immense amount of data has been collected in the Imnavait Creek watershed. Most of the major processes have been monitored continuously since 1985 [*Kane et al.*, 2004].

[31] Measurements collected from 2001 to 2003 are used in this study. Soil data from former studies complete the data collection. Sensors for air temperature, air pressure, wind speed, wind direction, relative humidity, radiation, soil temperature, and precipitation measure automatically. Except for the radiation measurements (March to September) the recording takes place throughout the year. All meteorological data used in this study are conducted at the Imnavait basin site. Liquid precipitation is measured using a tipping bucket rain gage equipped with a windshield. The threshold sensitivity of the tipping basket is 1 mm of rain, and the undercatch is estimated to be 5% (D. L. Kane, personal communication, 2007). The precipitation data used in this study have not been corrected to consider the undercatch. Stream discharge is estimated from stage data using a stage-discharge relationship. Discharge is measured from the beginning of the snowmelt until freeze-up. In July 2004, measurements were carried out at Imnavait Creek to obtain values for Manning's roughness parameter used in the modeling. These measurements were taken at two locations close to the flume station with both sections being several meters in length. An average value of 0.01 s/m$^{1/3}$ was determined, but is likely to be underestimated due to measurement restrictions [*Schramm*, 2005].

[32] The shallow soils consist of a layer of about 10 cm of organic material over 5–10 cm of partially decomposed organic matter mixed with silt which overlays the glacial till. Generally, there is a thicker organic layer in the valley bottom (~50 cm) than on the ridges (~10 cm). The soil parameters used in this study are based on a representative profile measured by *Hinzman et al.* [1991].

[33] Values for the annual active layer depth are based on Circumpolar Active Layer Monitoring (CALM) measurements (http://www.geography.uc.edu/~kenhinke/CALM/sites.html). The depth is measured each summer at the latest possible date prior to the annual freeze-up. The instrument used is a metal rod that is pushed vertically into the soil to the depth at which ice-bonded soil provides firm resistance. This determines the maximum depth of thaw (MDT). For Imnavait Creek, approximately 120 measurements are taken and averaged each year.

[34] The position of the water table used in this study is interpolated from measurements of volumetric soil moisture content made using time domain reflectometry sensors at seven depths within the soil profile at three sites located on the west facing slope of the watershed [*Overduin*, 2005].

[35] The SWE is measured late each spring just prior to snowmelt. To provide SWE data, snow depths are combined with pit studies to measure snow density, temperature, and hardness profile [*Reynolds and Tenhunen*, 1996]. The measurements are conducted along a valley transect, approximately in the middle of the basin. Each reported value is an average of at least 10 measurements [*Kane et al.*, 2001].

4.3. Calibration/Parameterization

[36] To simulate snowmelt, two methods are used to compare their ability to reproduce the snow pack ablation: the degree day method (model generated) and the energy balance method (calculated separately, as this method was not available at the time of this study). Concerning the degree day method, two parameters mainly determine the simulated snowmelt: the melt factor C_0 and the threshold value of the air temperature T_0. In this study a value of 2.3 mm/d °C for C_0, is found to produce the best results. T_0 is set to −1.2°C. When using the energy balance method for snowmelt (and later evapotranspiration), the average surface roughness length z_0 needs to be evaluated. In this study a constant value of 0.0013 m (0.02 m) for surface roughness length is used for the simulation of the melt period (evapotranspiration during summer). These values were determined by *Hinzman et al.* [1993]. Standard values are used for latent heat of fusion (3.34 10^6 J/kg), latent heat of vaporization (2.48 10^6 J/kg), water density (1000 kg/m^3), specific heat of air (1005.7 J/kg °C), density of air (1.2614 kg/m^3), and heat capacity of snow (2090 J/kg °C).

[37] Two methods are used in this study to calculate the amounts of water lost by evapotranspiration: the Priestley-Taylor method (model generated) and the energy balance method (calculated separately as this method was not available at the time of this study). For the Priestley-Taylor method the parameter α_{PT}, an empirical parameter, relates actual to equilibrium evaporation [*Priestley and Taylor*, 1972; *Rouse et al.*, 1977; *Mendez et al.*, 1998; *Kane et al.*, 1990]. In this study its calibration is based on the best alignment with the results obtained by the energy balance method, as this approach is physically based. Thus the best α_{PT} is determined to be 0.95. This value is used as an average for the entire watershed. For the thermal heat conductivity a value of 0.45 W/m °C was used that was determined through field measurements [*Hinzman et al.*, 1991].

[38] For the energy balance method, evapotranspiration is calculated as described by *Zhang et al.* [2000]. When this study was conducted, the energy balance methods (snow melt and evapotranspiration) were not incorporated into the model yet, and thus no spatially distributed variables could be used. This would have been possible for the degree day method and the Priestley-Taylor method, but was not done since the aim was to compare these methods to the results of the energy balance approach.

[39] The assignment of soil parameters to the horizontal soil layers (see Table 1) is based on studies by *Hinzman et al.* [1991] and the application of ARHYTHM to the same study site by *Zhang et al.* [2000]. When this study was conducted, a physically based representation of the active layer thawing process was not yet available. Instead, input files with changing hydraulic conductivities are used to

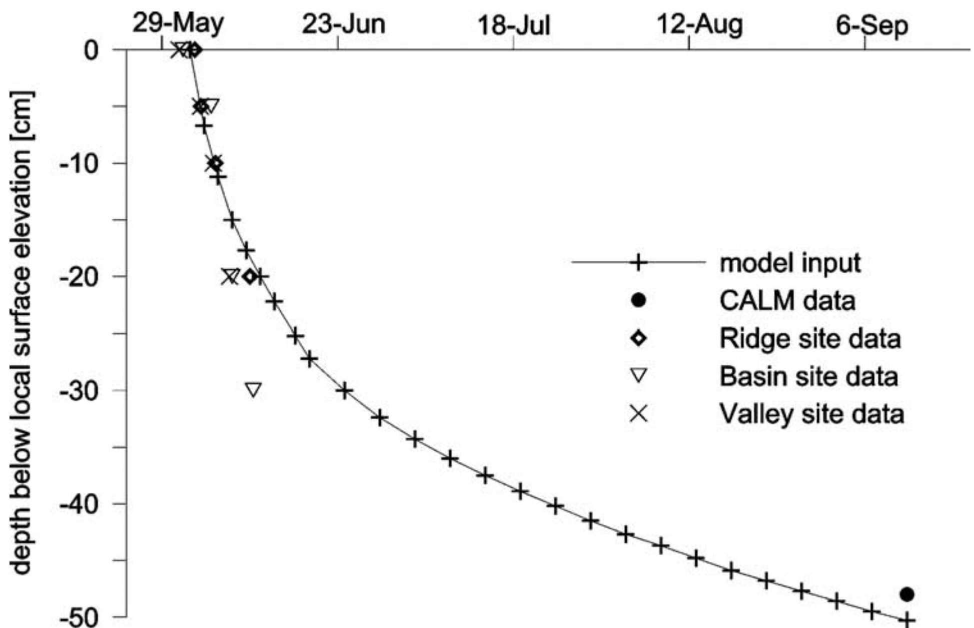

Figure 3. Thaw depth of the active layer 2001 used as a model input (α_{TD} = 0.068 during snow melt period 25 May to 14 June; α_{TD} = 0.032 during summer period 15 June to 13 September), determined from soil temperature measurements at the ridge, basin, and valley sites and from Circumpolar Active Layer Monitoring (CALM) grid measurements (average value).

account for the thawing of the soil. The soil is divided into layers of 10 cm, down to the maximum depth of thaw (MDT). During the course of the summer the thawing of the soil progresses and hydraulic conductivities are gradually (layer by layer) changed from frozen to unfrozen. The gradient controlling how the thaw depth evolves with time is determined by the α_{TD} value. The α_{TD} value is calibrated such that (1) during the initial thawing the input files match soil temperature recordings, and (2) at the end of the season the MDT matches the CALM grid measurements. Figure 3 shows the evolution of a gradually thawing active layer when used as a model input for 2001 and corresponding values obtained from measurements.

[40] When this study was conducted, the model did not allow the use of spatially distributed hydraulic conductivities and the thawing of the soil representing conductivities at the same time. In the case of a whole summer runoff simulation the thawing of the soil is an important factor and cannot be neglected. Thus the simulations are done on spatially homogeneous soil parameters.

[41] In this model, overland flow occurs when the water table rises above the surface. It is assumed that all of the water from precipitation or snowmelt is instantaneously infiltrated, meaning that the percolation time from the surface to the water table is neglected. The water content in each element may change with each time step, and the total storage capacity of each element may also increase or decrease as the active layer thaws.

[42] The crucial factor in determining overland and channel flow is the roughness parameter in Manning's equation [*Zhang et al.*, 2000]. In this study the coefficient is subjected to calibration within the range of values obtained from field measurements and literature [*Maidment*, 1992; *Emmett*, 1970]. For channel flow the channel bed width must be specified as well. Table 2 contains the corresponding values for each stream channel order.

5. Results

5.1. Water Balances 2001–2003

[43] The years 2001 to 2003 differ considerably in terms of hydrological and meteorological components. For the water balances (Figure 4), measured data are used for the rain, snow, and discharge components. Evapotranspiration is calculated with the energy balance method. The storage equals the residual term of the input (rain and snow) minus the output (discharge and evapotranspiration). Thus the storage term also includes the sum of errors caused by measurement uncertainties.

[44] In 2001 to 2003 the mean annual precipitation amounts to 337 mm, 520 mm, and 479 mm, respectively. Runoff accounts for 54%, 60%, and 67% of the water budget. The total amount of evapotranspiration is 48%, 42%, and 28% of the water budget. In each year the winter snow pack is a major source that adds water to the system. For the years of this study it accounts to 33–41% of the total amount of water added. A remarkable snow fall of 126 mm occurred in August 2002. The storage term,

Table 2. Overland and Channel Flow Parameters Used as Model Input[a]

	Manning's Roughness Parameter, s/m$^{1/3}$	Channel Bed Width, cm
Overland flow	0.30	-
Water tracks	0.15	5
Stream order 2	0.10	15
Stream order 1	0.07	40

[a]Data determined by field measurements and calibration.

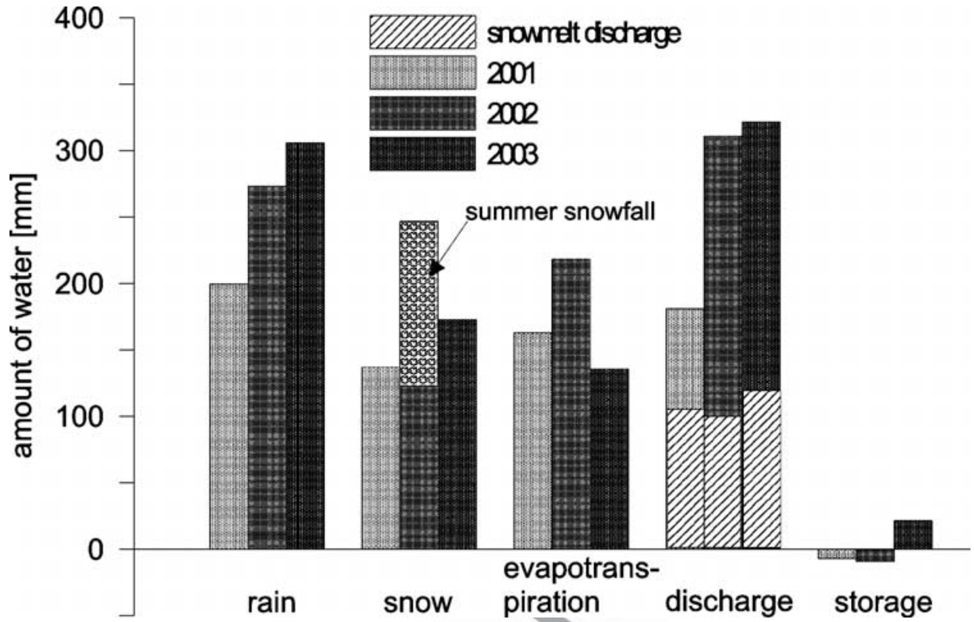

Figure 4. Water balance components for 2001–2003. Rain, snow, and discharge are based on measured data. Evapotranspiration is calculated using the energy balance method.

calculated as the residual term, shows little differences from year to year. Whereas in 2001 and 2002 the change in storage is slightly negative, there is a gain of 21 mm at the end of 2003.

[45] In the Imnavait watershed, 2001 represents an average year in most hydrologic components, whereas 2002 and 2003 show special characteristics that differ from mean values. 2003 is a wet year with continuously high precipitation, little evapotranspiration, high discharge, and a gain in soil moisture. Conversely, 2002 is characterized by the unusual summer snow fall and a high amount of evapotranspiration.

[46] Figure 5 shows the measured cumulative discharges of all years from the beginning of snowmelt until freeze-up, revealing distinct differences each year. The early onset of snowmelt in 2002 causes a considerably earlier start of discharge. Whereas in 2001 and 2003 the melt discharge is the highest discharge of the year, the peak discharge in 2002 originates from a snow/rain event in late summer.

[47] The influence of the antecedent soil moisture conditions on the runoff signal has been stated in section 2. This role is evident in each year of this study. For example, in 2002 the highest storm event of 9.3 mm/h recorded at 21 July results in a barely noticeable rise in runoff, after a

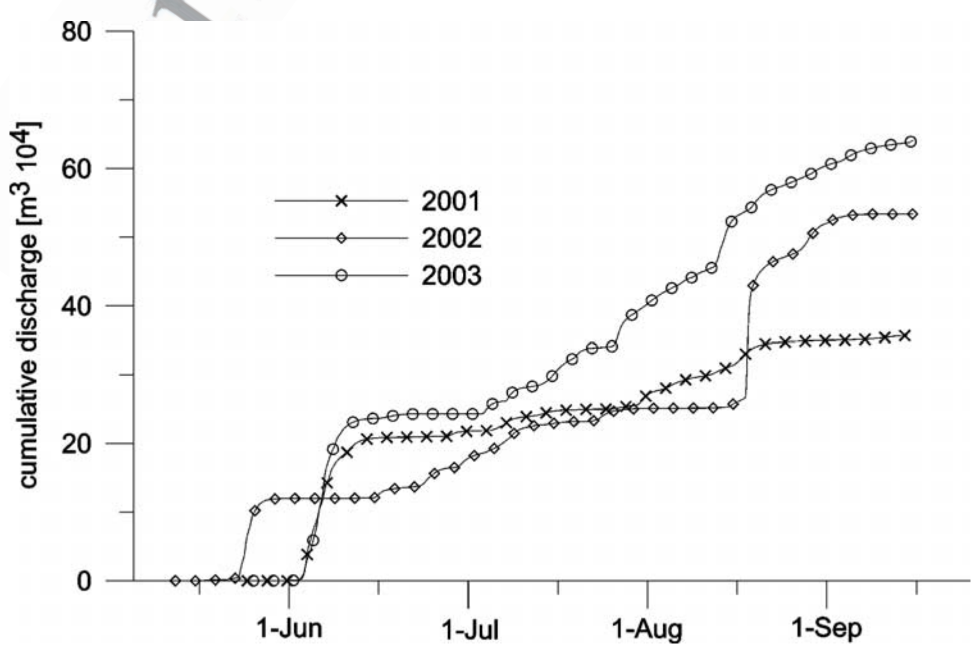

Figure 5. Measured cumulative discharges at Imnavait Flume station 2001–2003.

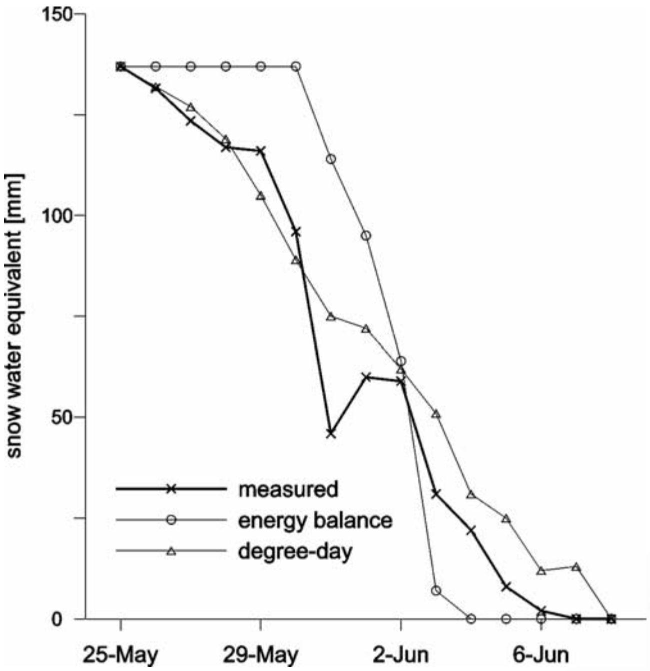

Figure 6. Measured and simulated snow ablation 2001.

7-h delay. Instead, a following rain event of 7 mm/h the next day generates a rise in discharge that exceeds the previous one by 3 times in peak and total amount. Also, the highest discharge on record with about 3.7 m^3/s is generated by a precipitation of 6 mm/h about 5 h earlier. In the first case a dry period of 7 d preceded the heavy rain event, whereas in the last two cases, precipitation was recorded previously.

[48] The discharge recorded at the end of the summer season 2003 shows an interesting feature not uncommon in arctic environments: At the time where the last peak occurs, freeze-up has already started, and surface temperatures show negative values for approximately 6 d. In addition, the last rain event that could have generated runoff is recorded 7 d prior to the peak in discharge. An explanation (R. E. Gieck, personal communication, 2004) for the occurring runoff could be that frazil ice and snow in the channel had blocked the outflow of one of the ponds upstream. When the ice dam broke, a small flood surge passed through the flume.

5.2. Modeling Results

5.2.1. Snowmelt

[49] In 2001 the snow pack ablated within 13 d. The initial SWE is obtained from snow survey measurements done prior to ablation. An average value is used for the entire watershed.

[50] Two methods, the degree day method (SM-DD) and the energy balance approach (SM-EB), are used to determine the snow pack ablation. SM-DD is used in the model simulation, whereas SM-EB is calculated separately. Figure 6 shows the simulated and the measured ablation curves for 2001. SM-DD achieves a better congruence than the energy balance method. Using the energy balance method, the onset of melt is delayed by 5 d, but completed earlier than measured. In the degree day method, the onset of snowmelt coincides exactly with the real onset, but the end of snowmelt is delayed.

[51] The discrepancy in congruence of the simulation and the recording could partly be due to the fact that field measurements are made daily in the morning, whereas both melt algorithms operate at hourly time steps. In addition, the pronounced spatial variability of the snow pack was stated previously, and other studies emphasize that the consideration of snow cover heterogeneity over complex arctic terrain provides a better representation of the end-of-winter snow water equivalent and an improved simulation of the timing and amount of water discharge due to snowmelt.

5.2.2. Discharge

[52] Measured versus simulated hydrographs for the year 2001, and the corresponding cumulative discharges, are depicted in Figure 7. It should be noted that because of the model configuration the simulation is split into snowmelt and summer period. The initial water table at the beginning of the summer simulation is set to the simulated height of the water table at the end of the snowmelt period.

[53] The diurnal fluctuations during the melt period, reflecting the influence of daily snowmelt cycles, are obvious in both, measured and simulated hydrographs. The onset of simulated discharge after snow melt occurs 7 d earlier than the measured one. Whereas this difference to the measured hydrograph is obvious, the total volume of melt discharge is very close to reality. The deviation in onset occurs because an algorithm for snow damming has not been incorporated into the model. Snow, redistributed by wind, accumulates in water tracks and valley bottoms, where melt water collects. The water seeps through the snow until it reaches a degree of saturation where both snow and melt water start to move, cutting a channel through the snow pack. *Kane et al.* [1989] found from measurements in the Imnavait watershed that the reduction of the snow water equivalent reaches up to 80% before stream runoff starts.

[54] Another explanation for the discrepancy between modeled and measured hydrograph could be the spatial variability of the snow pack. In this study an average value for initial SWE is used as an input, whereas in reality the variability of snow distribution with topography is pronounced [*Kane et al.*, 1991b; *Hinzman et al.*, 1996].

[55] During the summer runoff period the predicted cumulative discharge agrees well with the measured discharge volume. The simulated hydrograph caused by summer storm events shows some deviation from the recordings. For most rain events the simulated discharge leads measured data. Measured peak discharges are usually lower and have a longer recession time. The Nash-Sutcliffe coefficient for a weekly average is 0.64. An explanation for this discrepancy could be the beaded stream system, where small ponds act as reservoirs and store water intermediately, resulting in an attenuated hydrograph signal.

[56] Results indicate that the model performs well in the quantitative reproduction of the streamflow processes, but could be refined further in the timing of small-scale, short-term processes (see section 5.3).

5.2.3. Evapotranspiration

[57] Cumulative evapotranspiration and daily evapotranspiration rates for 2001 are shown in Figure 8. Evapotranspiration is only determined during the summer season. Priestley-Taylor (ET-PT) values are calculated by the

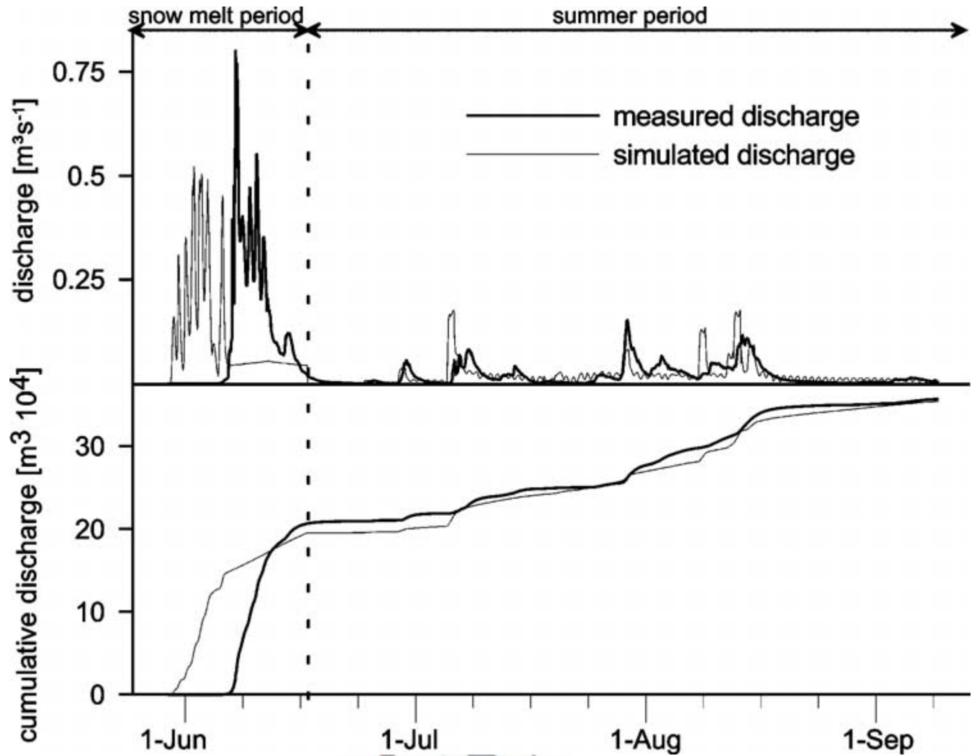

Figure 7. Measured and simulated discharge 2001.

model, whereas energy balance (ET-EB) calculations are done externally.

[58] In the total amount, ET-PT agrees well with the results of ET-EB. Figure 8 also illustrates the differences between ET-PT and ET-EB. Whereas fluctuations are pronounced in ET-EB, and fluxes are occasionally directed downward, ET-PT shows a steady rise without major fluctuations. This is due to the fact that both methods differ in the representation of the ventilation term, including the deficit in saturation and the wind component. ET-EB obtains this term from measurements, whereas in ET-PT this term is replaced by a constant. The ET-EB calculation shows the highest flux rates in early summer when both energy and water are relatively abundant.

5.2.4. Water Table

[59] Simulation results are compared with the measured water table height during summer 2003 at a water track site within the watershed (Figure 9). The year 2003 was chosen for this simulation, because measurements were available only for this period. Qualitatively, the simulation shows the

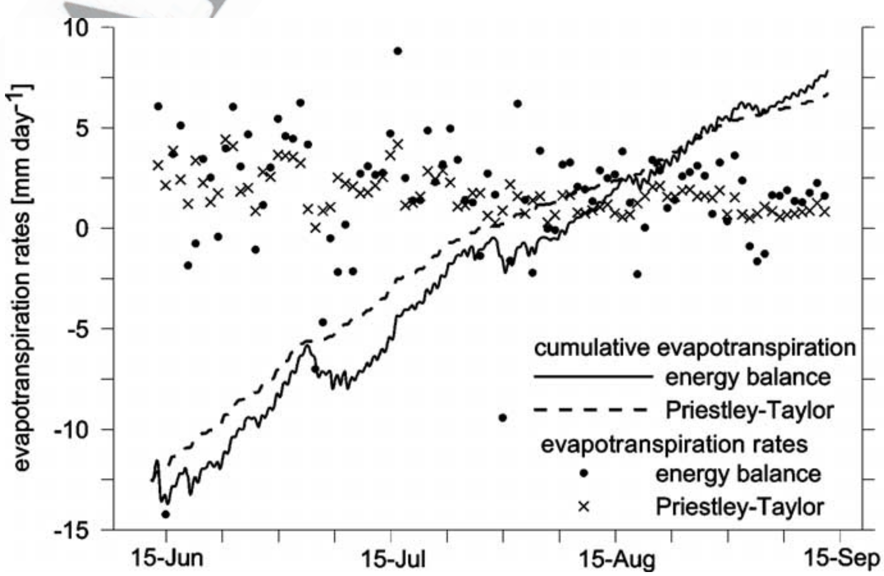

Figure 8. Cumulative hourly evapotranspiration 2001 and daily evapotranspiration rates 2001. The α_{PT} = 0.95 in the Priestley-Taylor calculation.

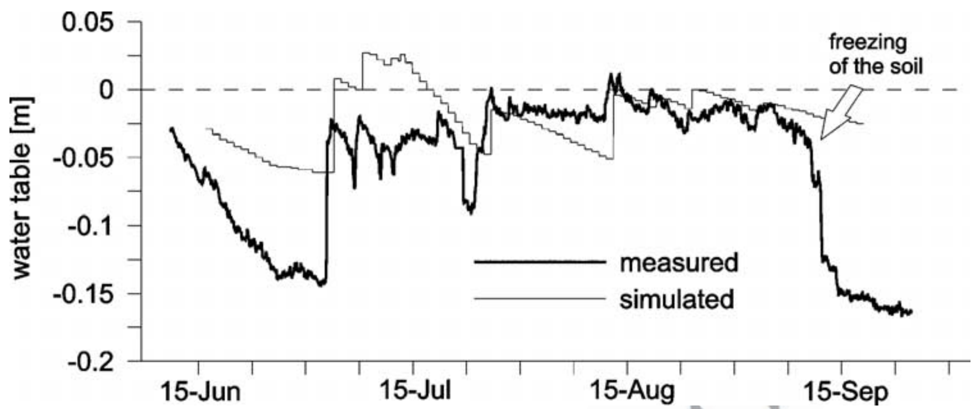

Figure 9. Comparison of simulated and measured water tables during the summer 2003 at a water track. Unit is water table (meters) relative to the local surface elevation. Refreezing of the soil results in the rapid decline of the measured water table in September.

same trends as the measurement. The sudden rises in the simulated water table are due to the instantaneous infiltration routine, where water percolation through the soil is neglected. The rapid decline in measured water table in September is caused by freezing of the soil. This process is not considered in the model simulations.

5.3. Model Sensitivity Toward Change in Parameters

[60] Figure 10a gives evidence of the influence of the MDT on total discharge. The importance of MDT is twofold: First, MDT has (in the current state of TopoFlow) to be given as input and thus underlies the uncertainties of measurements. For example, *Boike et al.* [1998] found that ground thaw depths determined using the probe method deviated considerably from the thaw depths determined by soil temperatures during the period when the active layer was dry. This is explained by a greater case of penetration of the frost probe when the active layer is saturated. Second, a simulation with increased MDT can reveal the runoff response to an increased melting of ground ice. In this study an increased MDT of 70 cm (compared to the normal case of 50 cm) is used in the summer simulation 2001.

[61] Figure 10b gives evidence of the importance of the initial water table height. It should be noted that only the

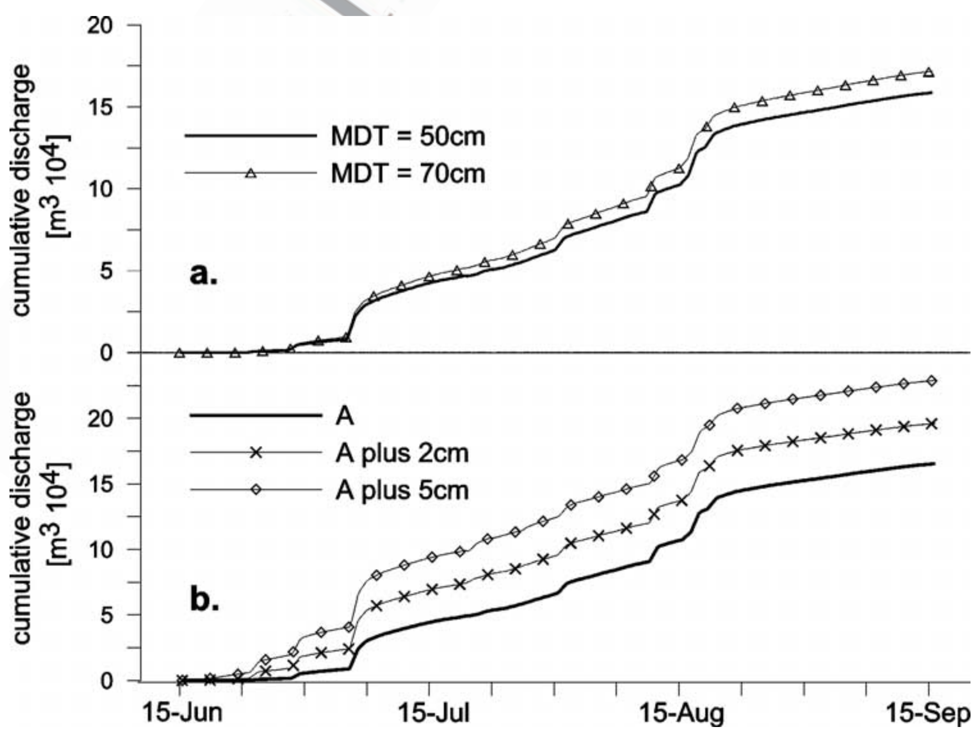

Figure 10. (a) Simulated discharge 2001 using different active layer depths. Maximum depth of thaw (MDT) is 50 cm in the normal 2001 simulation and was lowered to 70 cm for sensitivity studies. (b) Simulated discharge 2001 using different initial water table heights. Case A represents the normal water table height of the 2001 simulation. In the other simulations the water table height was raised by 2 cm and 5 cm, respectively.

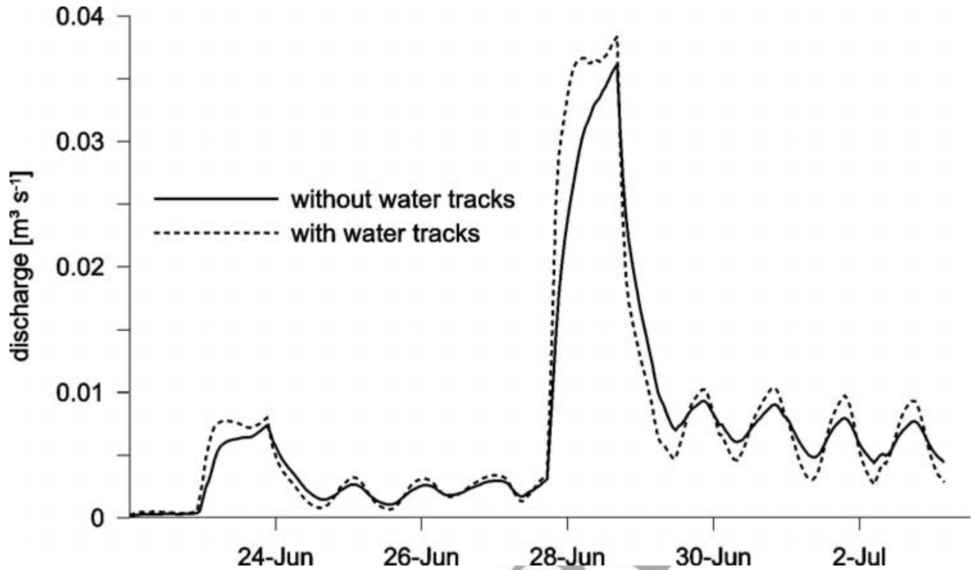

Figure 11. Simulated discharge 2001 illustrating the influence of water tracks. Solid line shows the normal 2001 simulation. Dashed line represents the discharge in a simulation where the water tracks were leveled to the adjacent surface elevation.

initial state of the water table height is given as an input, whereas its further evolution is calculated by the model. Here a small increase of 2 cm (5 cm), compared to the normal water table height (case A) causes an increase of 19% (38%) in the total amount of discharge. On the one hand, the influence of the antecedent soil water content on total discharge is characteristic for arctic watersheds where subsurface processes are limited to the shallow active layer. On the other hand, one should be aware of this sensitivity when calibrating the model.

[62] Figure 11 shows simulated hydrographs where the effect of water tracks (described in section 2) on the hydrograph is tested. The first simulation is based on the channel network depicted in Figure 2, whereas in the second simulation the water tracks are leveled to the adjacent surface elevation. The simulation indicates that the existence of water tracks accelerates runoff and leads to higher amplitudes in the hydrograph than would be present without them. Including water tracks improves the simulation result when compared to the measured hydrograph [*Schramm*, 2005]. Concerning the impact of soil parameters on subsurface flow, model studies reveal that the MDT has the highest influence, followed by the porosity and the hydraulic conductivity [*Schramm*, 2005].

5.4. Sensitivity of the Hydrological System Toward Changes in Climate Conditions

[63] As arctic temperatures and precipitation increase, there remains uncertainty on how the additional input of freshwater will be partitioned into streamflow and evapotranspiration. The interactions are further complicated by a contribution of melted ground ice to base flow when increasing temperatures deepen the active layer during summer. An open question is whether a change in climate will lead to a drying of the soil or to wetter conditions.

[64] Global and regional climate models predict different changes for the future climate state of the arctic depending on the warming scenario as well as on model performance. Regardless of the unanswered question, which scenario is the most likely one, changes on the hydrology can be investigated by presuming various conditions and using those as an input to model simulations. This was done in this study for different climate change scenarios (see Table 3) that include a change in three parameters: (1) the summer temperature, (2) the summer precipitation, and (3) the maximum depth of thaw.

[65] 2001 is used as the reference year, i.e., all changes of the above-mentioned parameters are relative to the observed climate conditions in 2001. Thus changes in the output, such as simulated runoff and evapotranspiration, can be compared to the 2001 simulation based on the real data set. Simulations were executed only for the summer season, lasting from 15 June until 13 September. The change in precipitation was distributed equally over the summer season, sustaining the range between minimum and maximum precipitation rates. For simplification the assumption of a 10 cm (20 cm) deepening of MDT by a 2°C warming was based on a study by *Kane et al.* [1991a]. The authors determine that a gradual but steady warming of 2°C would lead to a deepening of 10 cm (20 cm) after 20 a (45 a).

[66] Results indicate that a warming of 2°C without additional precipitation results in a higher R/P and ET/P ratio [*Schramm*, 2005]. Here the increase in runoff is

Table 3. Climate Change Scenarios A, C, and E and Their Changes in Mean Summer Temperature, Precipitation, and Maximum Depth of Thaw, Relative to the Observed Conditions in 2001

Scenario	A 10	A 20	C 10	C 20	E 10	E 20
Temperature, °C	+2	+2	+2	+2	+2	+2
Precipitation, %	-	-	+8	+8	−10	−10
Maximum depth of thaw, cm	58	68	58	68	58	68
Change in storage compared to 2001, mm	−7.8	−10.7	−3.5	−6.2	−7.9	−10.8

generated by a contribution of ground ice melted due to a deeper thaw depth. Runoff is significantly higher in the scenarios where an increase of precipitation is superimposed over the warming. The opposite accounts for the scenarios where precipitation input is decreased. All scenarios indicate an increased loss in storage compared to the reference amount in 2001, ranging from -3.5 to -10.8 mm. This indicates that the enhanced evapotranspiration overwhelms the increase in precipitation and results in a drying of the soil.

6. Conclusions

[67] This study presented the application of the hydrological model TopoFlow to Imnavait Creek, Alaska. Results indicate that the model is an excellent tool for simulating the overall water and energy balances of an arctic watershed. The model performs quantitatively well, with measured and simulated discharges being in a good agreement. The different components of the water cycle, i.e., evapotranspiration, snow melt, infiltration, and runoff, are well represented in the model, revealing that the model is able to handle the seasonal change in meteorological conditions. Some refinements are possible in the qualitative reproduction of some subprocesses: The onset of simulated snowmelt discharge occurs distinctly earlier than the measured discharge (7 d). This difference is in part due to the process of snow damming, which is not understood well enough to be incorporated into the model. Furthermore, the simulated summer hydrograph shows deviations from the recordings: Simulated discharge often leads site data; measured peak discharges are usually lower and have a longer recession time. This reveals that the model could be further refined in the small-scale, short-term reproduction of storage-related processes. Those can be attributed to the following facts: (1) The channel grid used in the simulation does not consider the ponds of the beaded stream system; (2) the spatial variability of the active layer depth is not represented in the simulation; and (3) the instantaneous infiltration used in the modeling simplifies the complex soil moisture distribution on short-term scales. Finally, simulation results could possibly be improved by spatially distributing several input variables (now possible in the model), such as snow depth, α_{PT} of the Priestley-Taylor method, the meteorological variables, and the soil parameters during the thawing season.

[68] Sensitivity studies reveal that the model is highly sensitive to the initial height of the water table that is given as an input to start the simulation. Even though this sensitivity is realistic, it requires calibration which naturally includes a source of error, as measurements are usually not available in full detail.

[69] While various studies present projected climate changes in the arctic, there remains uncertainty of how these changes will impact the hydrological cycle, resulting in enhanced or diminished runoff and soil moisture. This in turn is likely to affect the biogeochemistry and/or ecology of these systems (e.g., via changes in heat and water fluxes, vegetation cover, etc.). It is possible and desirable to couple TopoFlow with other models, and the authors encourage this development. This study shows that TopoFlow is a powerful tool for answering the question of how climate change will affect the sensitive wetlands of arctic tundra.

Notation

C_0	degree day melt factor, mm/d °C.
ET/P	evapotranspiration to precipitation ratio, no units.
R/P	runoff to precipitation ratio, no units.
T_0	temperature of snow for isothermal conditions, °C.
α_{PT}	alpha parameter controlling the thaw depth, no units.
α_{TD}	alpha parameter for the Priestley-Taylor equation, no units.
z_0	surface roughness length, m.

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References

ACIA (2004), *Impacts of a Warming Arctic: Arctic Climate Impact Assessment*, 1042 pp., Cambridge Univ. Press, New York.

Berezovskaya, S., D. L. Kane, and L. D. Hinzman (2005), Snowmelt hydrology of a headwater arctic basin revisited, *Eos Trans. AGU*, *85*(47), Fall Meet. Suppl., Abstract C41A-0195.

Bergström, S. (1976), Development and application of a conceptual runoff model for Scandinaian catchments, *SMHI Rep. RHO-7*, Swed. Meteorol. and Hydrol. Instit., Norrköping, Sweden.

Boike, J., K. Roth, and P. P. Overduin (1998), Thermal and hydrologic dynamics of the active layer at a continuous permafrost site, *Water Resour. Res.*, *34*(3), 355–363.

Bolton, W. R. (2006), Dynamic modeling of the hydrologic processes in areas of discontinuous permafrost, Ph.D. thesis, 163 pp., Univ. of Alaska Fairbanks, Fairbanks, Alaska.

Bolton, W. R., D. L. Hinzman, and K. Yoshikawa (2000), Stream flow studies in a watershed underlain by discontinuous permafrost, in *Water Resources in Extreme Environments*, edited by D. L. Kane, pp. 31–36, Am. Water Resour. Assoc. Proc., Anchorage, Alaska.

Bowling, L. C., D. L. Kane, R. E. Gieck, L. D. Hinzman, and D. P. Lettenmaier (2003), The role of surface storage in a low-gradient arctic watershed, *Water Resour. Res.*, *39*(4), 1087, doi:10.1029/2002WR001466.

Church, M. (1974), Hydrology and permafrost with reference to northern North America, in *Permafrost Hydrology, Proceedings Workshop Seminar*, pp. 7–20, Can. Natl. Comm., Int. Hydrol. Decade, Environ. Canada, Ottawa, Ont., Canada.

Dingman, L. S. (1973), Effects of permafrost on stream characteristics in the discontinuous permafrost zone of central Alaska, in *Permafrost, North American Contribution to the Second International Conference, Washington D. C.*, pp. 447–453, Natl. Acad. of Sci., Washington, D. C.

Emmett, W. W. (1970), The hydraulics of overland flow on hillslopes, *U. S. Geol. Surv. Prof. Pap., 662-A*, 68 pp.

Hastings, S. J., S. A. Luchessa, W. C. Oechel, and J. D. Tenhunen (1989), Standing biomass and production in water drainages of the foothills of the Philip Smith Mountains, Alaska, *Holarct. Ecol.*, *12*, 304–311.

Hinzman, L. D., and D. L. Kane (1992), Potential response of an arctic watershed during a period of global warming, *J. Geophys. Res.*, *97*, 2811–2820.

Hinzman, L. D., J. D. Fox, and D. L. Kane (1990), Soil freezing in a subarctic deciduous forest, in *International Frozen Soil Symposium, 21–22 March 1990, Spokane, Washington*, edited by K. R. Cooley, *CRREL Spec. Rep. 90-1*, pp. 21–30, U.S. Army Cold. Reg. Res. and Eng. Lab., Hanover, N. H.

Hinzman, L. D., D. L. Kane, R. E. Gieck, and K. R. Everett (1991), Hydrologic and thermal properties of the active layer in the Alaskan arctic, *Cold Reg. Sci. Technol.*, *19*, 95–110.

Hinzman, L. D., D. Wendler, R. E. Gieck, and D. L. Kane (1993), Snowmelt at a small Alaskan arctic watershed: 1. Energy related processes, in *Proceedings of the 9th International Northern Research Basins Symposium/Workshop Canada 1992*, edited by T. D. Prowse, C. S. Ommanney, and K. Ulmer, vol. 1, *NHRI Symp. 10*, pp. 171–226, Natl. Hydrol. Res. Inst., Saskatoon, Saskatchewan, Canada.

Hinzman, L. D., D. L. Kane, and Z. Zhang (1995), A spatially distributed hydrologic model for arctic regions, *International GEWEX Workshop on Cold-Season/Region Hydrometeorology, Summary Report and Proceedings, 22–26 May 1995, Banff, Alberta, Canada*, *Int. GEWEX Proj. Off. Publ. Ser. 15*, pp. 236–239, Washington, D. C.

Hinzman, L. D., D. L. Kane, C. S. Benson, and K. R. Everett (1996), Energy balance and hydrological processes in an arctic watershed, in *Landscape Function and Disturbance in Arctic Tundra*, *Ecol. Stud.*, vol. 120, edited by J. F. Reynolds and J. D. Tenhunen, pp. 131–154, Springer, Berlin.

Hinzman, L. D., et al. (2005), Evidence and Implications of recent climate change in northern Alaska and other arctic regions, *Clim. Change*, *72*, 251–298.

Hinzman, L. D., R. W. Bolton, K. Petrone, J. Jones, and P. Adams (2006), Watershed hydrology and chemistry in the Alaskan boreal forest: The central role of permafrost, in *Alaska's Changing Boreal Forest*, edited by F. S. Chapin III et al., pp. 269–284, Oxford Univ. Press, New York.

Kane, D. L., and L. D. Hinzman (1988), Permafrost hydrology of a small arctic watershed, in *Fifth International Conference on Permafrost*, edited by K. Senneset et al., pp. 590–595, Tapir Publ., Trondheim, Norway.

Kane, D. L., L. D. Hinzman, C. S. Benson, and K. R. Everett (1989), Hydrology of Imnavait Creek, an arctic watershed, *Holarct. Ecol.*, *12*, 262–269.

Kane, D. L., R. E. Gieck, and L. D. Hinzman (1990), Evapotranspiration from a small Alaskan arctic watershed, *Nordic Hydrol.*, *21*, 253–272.

Kane, D. L., L. D. Hinzman, and J. Zarling (1991a), Thermal response of the active layer in a permafrost environment to climatic warming, *Cold Reg. Sci. Technol.*, *19*(2), 111–122.

Kane, D. L., L. D. Hinzman, C. S. Benson, and G. E. Liston (1991b), Snow hydrology of a headwater arctic basin: 1. Physical measurements and process studies, *Water Resour. Res.*, *27*(6), 1099–1109.

Kane, D. L., L. D. Hinzman, J. P. McNamara, Z. Zhang, and C. S. Benson (2000), An overview of a nested watershed study in arctic Alaska, *Nord. Hydrol.*, *31*(4/5), 245–266.

Kane, D. L., K. M. Hinkel, D. J. Goering, L. D. Hinzman, and S. I. Outcalt (2001), Non-conductive heat transfer associated with frozen soils, *Global Planet. Change*, *29*(3–4), 275–292.

Kane, D. L., R. E. Gieck, D. C. Kitover, L. D. Hinzman, J. P. McNamara, and D. Yang (2004), Hydrologic cycle on the North Slope of Alaska, in *Northern Research Basins Water Balance*, *IAHS Publ.*, *290*, 224–236.

Maidment, D. R. (Ed.) (1992), *Handbook of Hydrology*, 1400 pp., McGraw-Hill, New York.

McCarthy, J. J., O. Canziani, N. A. Leary, D. J. Dokken, and K. S. White (Eds.) (2001), *Climate Change 2001: Impacts, Adaptation, and Vulnerability*, 1032 pp., Cambridge Univ. Press., New York.

McNamara, J. P. (1997), A nested watershed study in the Kuparuk River basin, arctic Alaska: Streamflow, scaling, and drainage basin structure, Ph.D. diss., Univ. of Alaska, Fairbanks.

McNamara, J. P., D. L. Kane, and L. D. Hinzman (1998), An analysis of streamflow hydrology in the Kuparuk River basin, arctic Alaska: A nested watershed approach, *J. Hydrol.*, *206*, 39–57.

Mendez, J., L. D. Hinzman, and D. L. Kane (1998), Evapotranspiration from a wetland complex on the arctic coastal plain of Alaska, *Nord. Hydrol.*, *29*(4/5), 303–330.

O'Callaghan, J., and D. Mark (1984), The extraction of drainage networks from digital elevation data, *Comput. Vision Graphics Image Process.*, *28*(3), 323–344.

Osterkamp, T. E., and V. Romanovsky (1999), Evidence for warming and thawing of discontinuous permafrost in Alaska, *Permafrost Periglacial Processes*, *10*, 17–37.

Oswood, M. W., K. R. Everett, and D. M. Schell (1989), Some physical and chemical characteristics of an arctic beaded stream, *Holarct. Ecol.*, *12*, 290–295.

Overduin, P. P. (2005), The physical dynamics of patterned ground in the northern foothills of the Brooks Range, Alaska, Ph.D. thesis, 182 pp., Univ. of Alaska, Fairbanks, Fairbanks, Alaska.

Priestley, C. H. B., and R. J. Taylor (1972), On the assessment of surface heat flux and evaporation using large-scale parameters, *Mon. Weather Rev.*, *100*, 81–92.

Quinton, W. L., D. M. Gray, and P. Marsh (2000), Subsurface drainage from hummock-covered hillslopes in the arctic-tundra, *J. Hydrol.*, *237*(1–2), 113–125.

Reynolds, J., and J. D. Tenhunen (Eds.) (1996), *Landscape Function and Disturbance in Arctic Tundra*, *Ecol. Stud.*, vol. 120, edited by J. F. Reynolds and J. D. Tenhunen, 437 pp., Springer, Berlin.

Romanovsky, V., M. Burgess, S. Smith, K. Yoshikawa, and J. J. Brown (2002), Permafrost temperature records: Indicators of climate change, *EOS Trans. AGU*, *83*(50), 589–594.

Rouse, W. R., P. F. Mills, and R. B. Stewart (1977), Evaporation in high latitudes, *Water Resour. Res.*, *13*(6), 909–914.

Schramm, I. (2005), Hydrologic modeling of an arctic watershed, Alaska, M. S. thesis, 108 pp., Univ. of Potsdam, Germany.

Serreze, M. C., J. E. Walsh, F. S. Chapin III, T. Osterkamp, M. Dyurgerov, V. Romanovsky, W. C. Oechel, J. Morison, T. Zhang, and R. G. Barry (2000), Observational evidence of recent change in the northern high-latitude environment, *Clim. Change*, *46*, 159–207.

Slaughter, C. W., and D. L. Kane (1979), Hydrologic role of shallow organic soils in cold climates, paper presented at the Canadian Hydrology Symposium, Natl. Res. Counc. of Can., Ottawa, Canada.

Smith, L. C., Y. Sheng, G. M. McDonald, and L. D. Hinzman (2005), Disappearing arctic lakes, *Science*, *308*(5727), 1429.

Stieglitz, M., J. Hobbie, A. Giblin, and G. Kling (1999), Hydrologic modeling of an arctic tundra watershed: Toward pan-arctic predictions, *J. Geophys. Res.*, *104*, 507–518.

Vörösmarty, C. J., L. D. Hinzman, B. J. Peterson, D. H. Bromwich, L. C. Hamilton, J. Morison, V. E. Romanovsky, M. Sturm, and R. S. Webb (2001), *The Hydrologic Cycle and Its Role in the Arctic and Global Environmental Change: A Rationale and Strategy for Synthesis Study*, 84 pp., Arct. Res. Consortium of the U. S., Fairbanks, Alaska.

Walker, D. A., E. Binnian, B. M. Evans, N. D. Lederer, E. Nordstrand, and G. W. Petersen (1989), Terrain, vegetation and landscape evolution of the R4D research site, Brooks Range Foothills, Alaska, *Holarct. Ecol.*, *12*, 261–283.

Walsh, J. E., et al. (2005), Cryosphere and hydrology, in *Arctic Climate Impact Assessment*, chapter 6, pp. 181–242, Cambridge Univ. Press, London.

Woo, M. (1986), Permafrost hydrology in North America, *Atmos. Ocean*, *24*(3), 210–234.

Woo, M., and P. Steer (1983), Slope hydrology as influenced by thawing of the active layer, Resolute N.W.T., *Can. J. Earth Sci.*, *20*(6), 978–986.

Zhang, Z., D. L. Kane, and L. D. Hinzman (2000), Development and application of a spatially-distributed arctic hydrological and thermal process model (ARHYTHM), *Hydrol. Processes*, *14*, 1017–1044.

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