SENSITIVITY ANALYSIS OF DISCHARGE IN THE ARCTIC USA BASIN, EAST-EUROPEAN RUSSIA

SANDRA VAN DER LINDEN [1], TARMO VIRTANEN [2], NAUN OBERMAN [3] and PETER KUHRY [4]

[1] *Department of Physical Geography, Utrecht University, P.O. Box 80.115, 3508 Utrecht, The Netherlands*
E-mail: S.vanderLinden@geog.uu.nl
[2] *Finnish Forest Research Institute, Rovaniemi Research Station, Rovaniemi, Finland*
[3] *Territorial Centre for State Monitoring, Ministry of Natural Resources and Environmental Protection of the Komi Republic, Syktyvkar, Russia*
[4] *Arctic Centre, University of Lapland, Rovaniemi, Finland*

Abstract. The high sensitivity of the Arctic implies that impact of climate change and related environmental changes on river discharge can be considerable. Sensitivity of discharge to changes in precipitation, temperature, permafrost and vegetation, was studied in the Usa basin, Northeast-European Russia. For this purpose, a distributed hydrological model (RHINEFLOW) was adapted. Furthermore, the effect of climate change simulated by a GCM (HADCM2S750 integration) on runoff was assessed, including indirect effects of permafrost thawing and changes in vegetation distribution. The study shows that discharge in the Usa basin is highly sensitive to changes in precipitation and temperature. The effect of precipitation change is present throughout the year, while temperature changes affect discharge only in seasons when temperature fluctuates around the freezing point (April and October). Discharge is rather sensitive to changes in vegetation. Sensitivity to permafrost occurrence is high in winter, because infiltration and consequently base flow increases if permafrost melts. The effect of climate change simulated by the scenario on discharge was significant. Peak flow can both decrease (by 22%) and increase (by 19%) compared with present-day, depending on the amount of winter precipitation. Also, runoff peaks earlier in the season. These results can have implications for the magnitude and timing of the runoff peak, break-up and water-levels.

1. Introduction

In the arctic environment, the impacts of climate change predicted by GCM experiments can be substantial, as warming is simulated to be greater there than at more southern latitudes (IPCC, 2001). Climate warming can result in the poleward movement of the permafrost boundary (Anisimov and Nelson, 1996), northward migration of the tree line (Serreze et al., 2000), and a change in hydrological processes (Woo et al., 1992; Van Blarcum et al., 1995; Rouse et al., 1997). Changes in the Arctic can have global impacts, because several feedback mechanisms come into play (Koster, 1993; Bonan et al., 1995). For example, a change in the spatial and temporal distribution of snow and ice can result in a change in albedo. Furthermore, tundra soils are an important repository of carbon (Post et al., 1982);

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climate warming could change tundra ecosystems from a net carbon dioxide sink to a source (Oechel et al., 1993).

A change in temperature and precipitation due to climate warming can affect evaporation, infiltration rates, soil moisture conditions, groundwater recharge and consequently timing and magnitude of runoff (Woo et al., 1992; Van Blarcum et al., 1995; Rouse et al., 1997). However, hydrological processes are not only affected by a change in climate variables but also by changes in vegetation cover and permafrost distribution. Evapotranspiration may intensify in newly forested regions if the tree line migrates northward, leaving less water available for runoff (Zhang et al., 2001). Permafrost zones will be displaced poleward due to a rise in temperature and the thickness of the active layer will increase (Anisimov and Nelson, 1996; Nelson et al., 2001).This can result in an increase in infiltration rates and subsurface storage, and consequently a decrease in surface runoff (Woo et al., 1992; Kuchment, 2000). The potential changes in precipitation, temperature, permafrost and vegetation are expected not only to affect runoff in the Arctic, but on a global scale, freshwater inflow from the Russian Arctic into the Arctic Ocean can modify global circulation patterns in the oceans (Bareiss et al., 1999; Serreze et al., 2000). The effects of climate change on discharge in Arctic areas remain poorly examined. Considering the potential local and global consequences, studies to the hydrological impacts of climate and environmental change in the Arctic are of utmost importance.

This paper represents a contribution to the TUNDRA project (Tundra Degradation in the Russian Arctic) that studies Arctic feedback processes to the global climate system. The goal of this study was twofold: (i) to qualify the hydrological sensitivity of an Arctic basin to climate change and related environmental changes, such as vegetation redistribution and permafrost thawing, and (ii) to investigate the hydrological impacts of climate change derived from a GCM experiment. In the sensitivity study, discharge was simulated using a distributed water balance model for various hypothetical meteorological (precipitation and temperature) and environmental (permafrost and vegetation) changes. The temperature changes predicted by the GCM experiments were used to determine the effect of climate change on permafrost and vegetation. Together with the climate variables these were used in the model to simulate discharge. The study area is the Usa basin in the European Russian Arctic that is unique in continental Europe for having extensive lowland tundra and permafrost. The region is particularly suitable to study the effects of global change in the Arctic because it includes major features such as the Arctic and alpine (in the Ural Mountains) tree lines and the limits of continuous and discontinuous permafrost, which are all very sensitive to climate changes.

2. Area Characteristics

The Usa basin is located in East-European Russia (56°–66° E, 64°–68° N), Figure 1. The catchment (93,000 km^2) is bordered by the Ural Mountains in the east. At the west side of the catchment, the Usa river discharges into the Pechora river. The Ural Mountains comprise approximately 15% of the area, where elevation ranges from 300 to 1800 m. The remaining part of the basin has an elevation between 40 and 300 m, mostly below 200 m. Mean annual temperature ranges from –3 °C in the south to –7 °C in the northernmost regions. In winter, temperature can be as low as –55 °C, while summer temperatures up to 35 °C are measured. In the Ural Mountains, mean annual precipitation is 950 mm. In the lowland areas, mean annual precipitation ranges from about 400 to 800 mm (Taskaev, 1997; Christensen and Kuhry, 2000). Maximum precipitation occurs in summer with values up to 100 mm/month in August (Figure 2). A hydrograph of the Usa river at the Adzva station is shown in Figure 2. Discharge is characterised by a typical (sub-)arctic flow regime (Church, 1974). In autumn and winter (October until April), a minor base flow is sustained in areas where discontinuous permafrost allows some discharge of groundwater. Base flow is absent in flat areas with continuous permafrost, where winter temperature is extremely low. A major runoff peak occurs in May or June due to snowmelt, as soon as temperature rises above the freezing point. After the snowmelt period, runoff decreases to less extreme values with occasional runoff peaks due to heavy rainstorms in summer (July until September).

The northern part of the Usa basin is covered with treeless tundra vegetation and peat plateau mires. The upland areas are vegetated by shrub tundra vegetation with a well-developed lichen and/or moss layer. Willow-dominated, often paludified, vegetation occurs in depressions and river valleys. The central part of the Usa basin consists of a mosaic of tundra and northern taiga forests. The southern part belongs to the northern taiga forest zone. Large open mires with isolated palsas are common in the lowlands of the taiga zone. Forest stands in lowland areas mainly consist of mixed forests dominated by spruce (*Picea obovata* Ledeb.), while common white birch (*Betula pubescens* Ehrh.) is the dominant deciduous tree species. Pines (*Pinus sylvestris* L.) are rare and are found some tens of kilometres south of the spruce tree line, mostly around open mires. Apart from the surrounding areas of a few towns and industrial areas, there are no forest clearings. In the alpine taiga zone, forest consists of spruce, larch (*Larix Sibirica* Ledeb.), Siberian fir (*Abies Sibirica* Ledeb.) and birch. Different types of alpine shrub and grass tundra including luxurious meadows are found above the alpine tree line. The steepest slopes and highest mountain tops are mainly stony and barren.

In the flat parts of the Usa basin, permafrost occurs as isolated patches in the south, gradually extending to continuous permafrost in the north. In the Ural mountains discontinuous and continuous permafrost occur. Permafrost temperatures range from –4.5 °C in the northern lowland area to just below freezing point in the southern region. In the Ural mountains, permafrost temperatures can decrease

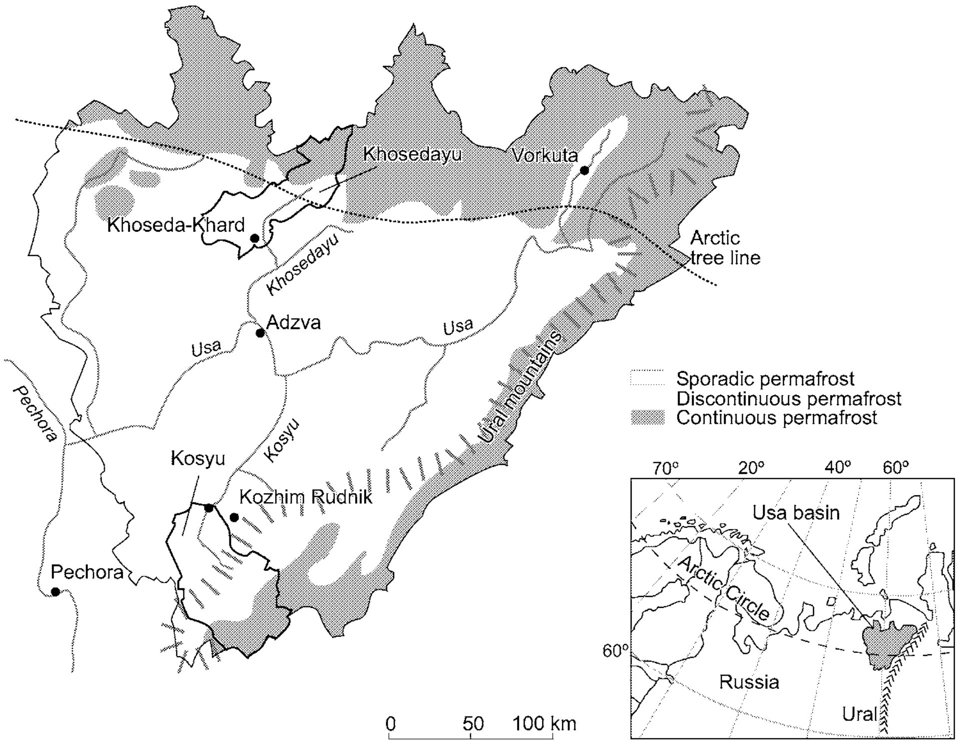

Figure 1. The Usa basin and two subcatchments (Khosedayu and Kosyu) with the hydrological and meteorological stations.

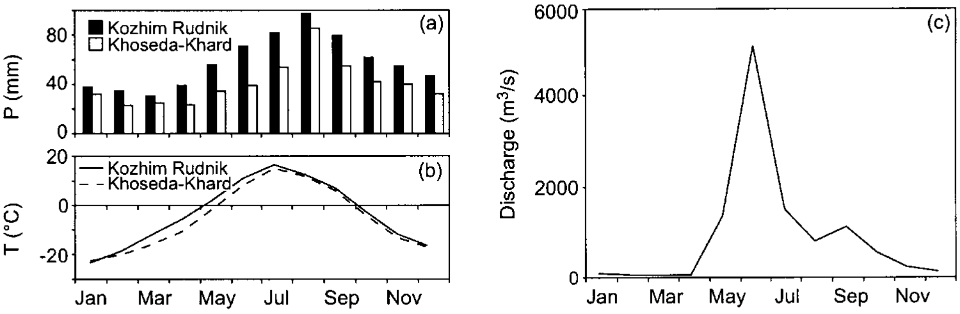

Figure 2. Mean monthly precipitation (a) and temperature (b) at Kozhim Rudnik and Khoseda-Khard and runoff (c) of the Usa at the Adzva station.

to –7 °C, due to the lack of a snow cover as a result of low winter precipitation. The base of the permafrost is found at depths between 10 and 700 m below surface.

Two subcatchments representative for a specific part of the Usa basin were studied in more detail (Figure 1). The Khosedayu catchment includes large areas of lowland tundra, which is characterised by extensive peat plateau mires. Some isolated spruce stands are found in the river valleys and in the well-drained uplands in the south. The area has little relief and most of the area is underlain by sporadic and discontinuous permafrost (10–50% coverage) of which the temperature is between –1 and –2 °C. Soils consist mainly of loam and clay with a high volumetric ice content. The depth of the permafrost base is about 400–500 m. Mean annual air temperature is –6.5 °C and annual precipitation at the nearby Khoseda-Khard climate station is 430 mm. The Kosyu catchment is located in the northern boreal taiga zone where spruce dominated forests and extensive open peatlands prevail. The headwaters of the catchment are located in the Ural Mountains (elevation up to 1900 m). Permafrost is restricted to isolated patches in the lowlands, although more extensive permafrost occurs in the Ural Mountains. Mean annual temperature is –3 °C and annual precipitation is 600 mm at the nearby station Kozhim Rudnik.

3. Data and Description of the Model

3.1. HYDROMETEOROLOGICAL DATA

Meteorological data were obtained from the Komi Republican Center for Hydrometeorology and Environmental Monitoring, Syktyvkar, Russia. Series of monthly precipitation and temperature were available for 13 meteorological stations in the Usa basin. All stations are located in the lowland area or in the foothills of the Ural Mountains, while no stations are present in the Ural Mountains where high precipitation and extreme temperature ranges occur. Time series of observed monthly temperature and precipitation cover the period 1950–1987. In the early part of the series, records for only a few months each year were available for most stations. After 1970, records are almost complete.

Temperature and precipitation data were interpolated over the basin using simple Thiessen polygons. The temperature in a grid cell was obtained from the temperature of the representative Thiessen polygon and was corrected for the altitude of that cell, assuming a lapse rate of 0.6 °C per 100 m. The lack of observed precipitation data in the Ural mountains would result in an underestimation of areal precipitation. Therefore, simulated precipitation data was used from the regional climate model HIRHAM4 (Christensen et al., 1996) to estimate the spatial pattern in precipitation over the basin. HIRHAM4 overestimates areal precipitation, but the spatial distribution of simulated precipitation is physically consistent (Christensen and Kuhry, 2000). Therefore, the spatial distribution of precipitation obtained from HIRHAM4 was combined with observed precipitation data as described in Van der Linden and Christensen (2002). This method can be used to create precipitation time series even for periods with no climate model data available, as in this paper.

Also, discharge data were obtained from the Komi Republican Center for Hydrometeorology and Environmental Monitoring, Syktyvkar, Russia. Discharge data at the outlet of the Usa river are not available. The most downstream hydrological station with monthly discharge data is the Adzva hydrological station (catchment area is 66,080 km^2). Monthly discharge data for Kosyu (4030 km^2) are available from the Kosyu station and for Khosedayu (2560 km^2) from the Khoseda-Khard station (Figure 1). Discharge records represent the period 1950–1990.

3.2. EVAPOTRANSPIRATION DATA

Evapotranspiration equations that are commonly used in modelling either need input data that are not available for the Usa basin (e.g., Penman, 1948; Priestley and Taylor, 1972; Spittlehouse, 1989), or do not give reliable estimates in cold environments (e.g. Thornthwaite and Mather, 1957). Therefore, potential evapotranspiration was based on evapotranspiration simulated by HIRHAM4. The climatology at the monthly scale as simulated with HIRHAM4 was thoroughly evaluated by Christensen and Kuhry (2000). They found that the model was depicting the areal distribution realistically when compared to the limited amount of station data available in the vicinity of the Usa river catchment. This gives confidence in the spatial pattern of evapotranspiration predicted by HIRHAM4. However, due to the overestimation of precipitation by HIRHAM4, the model also overestimates evapotranspiration (cf. Hagemann et al., 2001). In this study, a comparison of simulated actual annual evapotranspiration (estimated using the HIRHAM4 potential evapotranspiration and soil moisture availability over a period of ten years) and annual evapotranspiration measured over 40 years in the Usa basin (documented in Taskaev, 1997) revealed that HIRHAM4 overestimates evapotranspiration by 20%. Therefore, it was considered necessary to reduce the potential evapotranspiration by 20%. The HIRHAM4 simulation was carried out for the period 1979–1993. Spatial resolution of the evapotranspiration output is approximately 16 km. Evapotranspiration data in each grid cell were summed for each month to create series of spatially distributed monthly potential evapotranspiration.

3.3. DIGITAL ELEVATION MODEL, VEGETATION CLASSIFICATION AND PERMAFROST MAP

A digital elevation map (DEM) of the Usa basin was derived from 1:200,000 digital topographic maps with drainage network, lakes and contour lines purchased from the State GIS Centre (GOSGISCentre), Moscow, Russia. From this data a DEM with a grid size of 1 km was constructed. Vegetation data was produced by classifying a Landsat 5 TM mosaic of the study area (pixel size 30 m). Training data on vegetation type was derived from ‘ground truth’ plots sampled during field work in the summers of 1998, 1999 and 2000. Twenty land cover classes were distinguished. The classified Landsat TM mosaic was converted to a digital map with grid cells of 1 km. A grid cell in the map was classified as forest, if more than 20% of the area in the mosaic was classified as forested, the other cells were classified as non-forested. A digital permafrost map including information on types and temperature of the permafrost in the Usa basin was developed by the stock company PolarUralGeologia, Vorkuta, and the Komi Science Centre, Syktyvkar, Russia.

3.4. HYDROLOGICAL MODEL

To study the sensitivity of the hydrological regime for changes in climate, vegetation and permafrost, a simple distributed hydrological model (USAFLOW) was used. USAFLOW is a GIS-based model that calculates the water balance on a monthly basis. Monthly time steps were used, because only monthly input data were available. This makes the calculation of the exact timing of the transition between rain and snowfall impossible, but other studies showed that this approach still allows to determine te essential characteristics of the discharge regime in large basins (Kwadijk, 1993; Van Deursen, 1995). The model was derived from RHINEFLOW, a model that was developed for large river basins and was applied successfully in the Rhine and Meuse basin, Europe, the Ganges-Brahmaputra basin, India and the Yangtze basin, China (Kwadijk, 1993; Van Deursen, 1995). USAFLOW uses the following input variables: monthly temperature, precipitation and evapotranspiration, and a Digital Elevation Map (DEM). The data are stored in a GIS using a grid of $1 * 1$ km^2. A resolution of $1 * 1$ km^2 was used, because it allows the use of the model in the smaller subbasins and it makes optimal use of the altitudinal information present in the DEM.

A flow diagram showing the sequence of steps carried out by the model is presented in Figure 3. The water balance is calculated for each grid cell on a monthly basis. Precipitation is treated as rain or snow depending on a critical temperature (0 °C). Snow is stored at the surface until snowmelt takes place. Snowmelt is calculated with the degree-day method using a critical temperature and a snowmelt factor. Both rainfall and snowmelt amounts are summed for each time step, and this water is directly available for evapotranspiration, infiltration and runoff. To estimate evapotranspiration for the period for which HIRHAM4 evapotranspiration data were not available, a method described by Gellens and Roulin (1998) was used (equation given in Figure 3). This method is based on an empirical relationship between temperature and evapotranspiration. Actual evapotranspiration is the minimum of available water and potential evapotranspiration, and is subtracted by the model from the water present at the surface. Using a defined separation coefficient, the soil water surplus (available water at surface minus evapotranspiration) is divided, in a part that is discharged directly and a part that percolates to the groundwater reservoir, where it remains stored for a longer period. A coefficient of 0 results in the absence of percolation to the groundwater storage, so that all water is discharged within the same month as direct runoff. Each time step part of the groundwater is discharged as delayed runoff. The amount of delayed runoff is estimated using a recession coefficient. Finally, discharge is generated at the catchment outlet by summing the excess water (direct and delayed runoff) of all upstream cells. This procedure is repeated for each monthly time step.

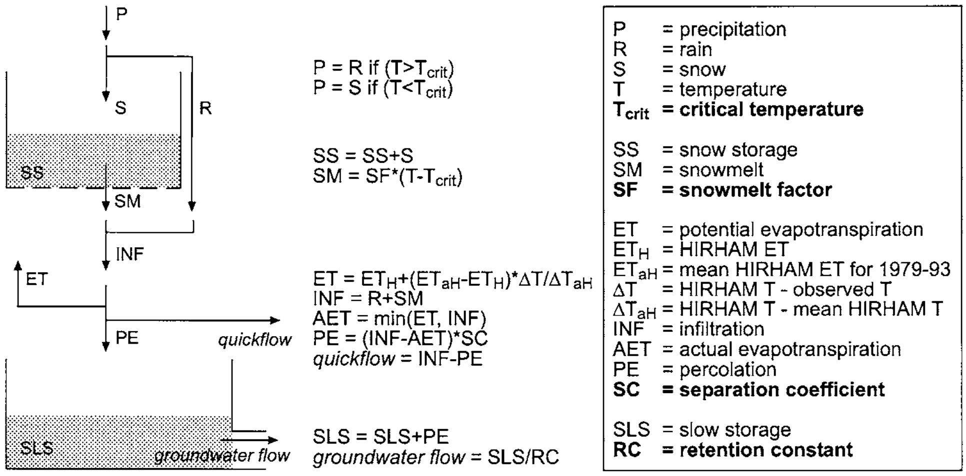

Figure 3. Flow diagram of the USAFLOW hydrological model (parameters in bold).

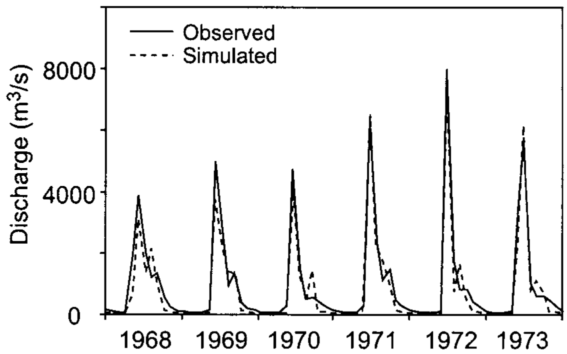

Figure 4. Observed and simulated discharge for 1968–1973 at the Adzva station (Usa river).

Snowmelt factor, separation and recession coefficients were estimated by model calibration on the basis of discharge data obtained at the Adzva hydrological station. First, the model was calibrated for a relative wet period (1980–1984) and subsequently, the model was validated for a relative dry period (1969–1973). Modelled discharge was in close agreement with observed discharge with an R^2 (Nash and Suthcliffe, 1970) of 0.9 for both periods. In Figure 4, model results for the validation period are shown together with observed discharge.

The values for the snowmelt factor (200 mm/month) and the recession coefficient (2) were kept constant over the entire basin and over time. In reality, the snowmelt factor varies as sun angle, cloudiness, and internal characteristics of the snow cover change. Within a year, the effect of these factors is limited, because a large amount of snow melts within one time step. However, under changed climate conditions, a change in cloudiness can affect the value for the snowmelt factor. This possible effect is not taken into account for this study, because no reliable data are available to quantify a change in cloudiness. The value for the recession coefficient depends mainly on the geohydrological properties of the catchment and these properties will not alter under changing climatic conditions (Kwadijk, 1993). Different values for the separation coefficient were established for three zones, the sporadic permafrost zone (0.3), discontinuous permafrost zone (0.2) and continuous permafrost zone (0.1). These values are in agreement with data of Bratsev (1982), who found that 90% of the available water at the surface contributes to direct runoff in the discontinuous permafrost regions. In southern regions where permafrost is absent, about 70% of the surface water is discharged directly. Thus, the separation coefficient is expected to increase when permafrost degrades due to climate warming. The separation coefficient also depends on slope, soil physical properties, land use and variations in soil moisture conditions over time. However, no information was available to quantify the effects of these properties.

4. Sensitivity Analysis

The sensitivity study in the Usa basin was done using the meteorological record of 1968–1973 as a reference. Discharge was simulated at the outlet of the Usa river to examine the sensitivity of discharge to changes in temperature, precipitation, areal extent of permafrost and vegetation in the Usa basin. The sensitivity of discharge in the Kosyu and Khosedayu subcatchments was examined as well.

To investigate the sensitivity to a change in meteorological variables, the present-day observed temperature and precipitation input (1968–1973) was changed. Temperature was reduced by 4 °C and increased by 2 and 4 °C. Precipitation was reduced by 20% and increased by 10% and 20%. These changes in temperature and precipitation are arbitrary but fall within the range of possible change (IPCC, 2001), except for the decrease in temperature. Subsequently, the sensitivity to changes in vegetation and permafrost was tested. Vegetation type partially influences the amount of evapotranspiration. Forests show higher evapotranspiration rates than lichen-covered tundra regions (Zhang et al., 2001). Therefore, sensitivity estimates for a change in vegetation were carried out by varying effective evapotranspiration. The evapotranspiration was increased by 10 and 20% and reduced by 20%. The changes in the evapotranspiration are hypothetical, but fall within ranges found at present, i.e., in the Usa basin evapotranspiration measured in the tundra-covered north (220 mm) is 30% lower than in the taiga-covered south (285 mm) (Taskaev, 1997). Permafrost affects the amount of water that can infiltrate into the soil, which is represented in the model by a separation coefficient. A higher percentage of area underlain by permafrost is represented by lower separation coefficient values and results in higher amounts of direct runoff. To study the sensitivity to a change in permafrost, the separation coefficient was decreased to 0 and increased by a factor 2 and 4. The changes in the separation coefficient are arbitrary, but fall within ranges found at present by Bratsev (1982).

Sensitivity of discharge was tested on an annual basis and for each month separately. A model simulation was carried out with present-day values of precipitation, temperature, evapotranspiration (vegetation) and the separation coefficient (permafrost). Then, the model was run using separately the changed climate and environmental variables and the relative change in discharge was calculated.

5. Scenario Study

5.1. CLIMATE SCENARIOS

The climate scenarios for the 2080s and 2230s used in this study were obtained from the HadCM2 Climate Change Experiments (Hadley Centre, Bracknell, U.K.) through the Climate Impacts LINK Project (Climatic Research Unit, University of East Anglia, Norwich, U.K.). The second generation Hadley Centre coupled ocean-atmosphere GCM is described in Mitchell et al. (1995) and Johns et al. (1997). The HadCM2 model uses a global grid of 96 longitude (every 3.75 degrees) $\times$ 73 latitude (every 2.5 degrees). At this resolution only five grid points are in or near the Usa Basin (67.50 N–56.25 E, 77 m; 67.50 N–60.00 E, 122 m; 67.50 N–63.75 E, 186 m; 65.00 N–56.25 E, 158 m; 65.00 N–60.00 E, 229 m), providing a very poor representation of the Urals mountains. The ability of GCMs to accurately represent Arctic climates is limited (e.g., Chen et al., 1995; Tao et al., 1996). A comparison of modelled climate at the five grid points with five nearby located weather stations in the study area for the period 1961–1990 shows that mean annual air temperature is underestimated on average by 1.6 °C , with maximum departures in summer (July temperature is especially under-estimated by 6.2 °C). Annual precipitation is over-estimated in the model by on average 18%, with greatest differences in the months March till May (up to 50% over-estimation).

The use of the HadCM2S750 stabilisation run provides an opportunity to model vegetation, permafrost and hydrology in the Usa Basin under transient climate change (2080s, indicated further as the 2080 scenario) and under equilibrium conditions reached by the 2230s (indicated further as the 2230 scenario). The climate changes predicted by HadCM2S750 are consistent with the results of other GCM experiments (Räisänen, 2000), but they are smaller than the changes obtained by the GCM runs using the newest IPCC (SRES) greenhouse gas concentrations and aerosol levels (Cubash et al., 2001; Johns et al., 2001).

The calculated anomalies indicated in Table I are the mean of 30-year averages for the five grid points between the 1961–1990 control run (HadCM2GGa1 integration) and the time intervals 2070–2099 and 2220–2249 derived from the 750 ppm stabilisation run (HadCM2S750 integration), which stabilises greenhouse gas concentrations from 2200 onwards. The use of a mean was considered acceptable because climatic gradients among the five grid points remain the same between the control and stabilisation runs.

Table I

30-Year mean monthly and annual climate change anomalies in mean temperature and total precipitation for 2070–2099 and 2220–2249 in the 750 ppm stabilisation run (HADCM2S750), compared to the 1961–1990 control run (HAD2CMGGA1)

	2070–2099		2220–2249	
	T (°C)	P (%)	T (°C)	P (%)
January	0.9	98	5.4	143
February	2.7	111	5.6	124
March	2.8	128	3.4	128
April	4.3	116	3.5	112
May	2.7	109	3.9	119
June	2.3	102	3.0	119
July	2.8	116	3.9	133
August	2.5	106	4.3	115
September	2.6	105	3.4	113
October	2.7	114	2.9	118
November	2.0	108	3.1	131
December	5.2	119	6.5	123
Annual	2.8	110	4.1	123

5.2. TREE LINE DYNAMICS

According to the vegetation data, the most important land cover types in the Usa basin are forests, shrub dominated tundra, and peatlands (24%, 27% and 30%, respectively). A potential impact of climate warming on the distribution of vegetation is a northward shift of the tree line. Logistic regression models on the basis of present climate were developed to estimate the location of the future tree line in the Usa basin. The model with the highest correlation between the distribution of forest and a combination of annual and monthly precipitation, evaporation, mean, minimum and maximum temperature, and degree-day sums was used. The assumption was made that grid cells dominated by peatlands, water bodies and bare areas will remain unchanged under future climate change. The best prediction of forest coverage was achieved with mean July temperature and minimum annual temperature (94.3% of the observations were predicted correctly). When the regression model is used with the 2080 scenario, the entire Usa basin falls within the range of forest growth, except for the highest mountain areas. However, due to time lags in seed dispersal and growth rates, it is not reasonable to assume that the entire area is already forested by the 2080s. However, by the 2230s+ (equilibrium situation) all soils that are suitable for forest growth will be forested.

5.3. PERMAFROST DYNAMICS

Permafrost occurrence will decrease due to increasing temperature. The assumption is made that the increase in temperature of the surficial permafrost is the same as the increase in air temperature. Monitoring studies in the Usa basin for the period between 1970 and 1995 indicate that this assumption is correct for soils with a high volumetric ice content and consequently a high heat conductivity (Oberman, 2001). Permafrost temperature in soils with a low ice content increases twice as slowly as the air temperature. However, it is believed that the duration of the period up to the 2080s is sufficiently long for the permafrost temperature to adapt. Furthermore, model results from Romanovsky (pers. comm.) indicate that the relatively warm permafrost close to Vorkuta (in the Usa basin) can degrade rapidly if temperature increases according to the HADCM2S750 scenario, with talik development to 5 m below surface by the 2080s. Consequently, discontinuous permafrost is replaced with thawed ground and sporadic permafrost.

5.4. HYDROLOGICAL CHANGES

Discharge was simulated for the 2080 and the 2230 scenarios, taking into account changes in climate, vegetation and permafrost simultaneously. The monthly changes in temperature and precipitation according to the climate scenario were added to the present-day temperature and precipitation to create time series of future temperature and precipitation. No change in vegetation was applied for the 2080 scenario. For the 2230 scenario the assumption was made that all areas presently occupied by tundra vegetation (excluding peatlands) will be forested. According to data presented by Zhang et al. (2001), evapotranspiration may increase by about 40% if non-forested vegetation is replaced with forest. An increase in forest cover over the entire Usa basin (from 25% to 51%), and in the Khosedayu (from 3 to 59%) and Kosyu basins (from 46 to 53%) results in an increase in evapotranspiration by 10%, 20% and 3%, respectively. By the 2080s, discontinuous permafrost will be replaced with thawed ground and sporadic permafrost. In the discontinuous permafrost zone, 90% of the surface water contributes to direct runoff, while in areas with thawed ground only 70% of the surface water is discharged directly (Bratsev, 1982). Therefore, the separation coefficient in the model was increased to 0.3 in areas where permafrost is expected to be disappeared in future.

6. Results of the Sensitivity Analysis

The discharge series simulated with varying precipitation for the Usa, Kosyu and Khosedayu river are given in Figure 5 (see also Table II). For all catchments, the

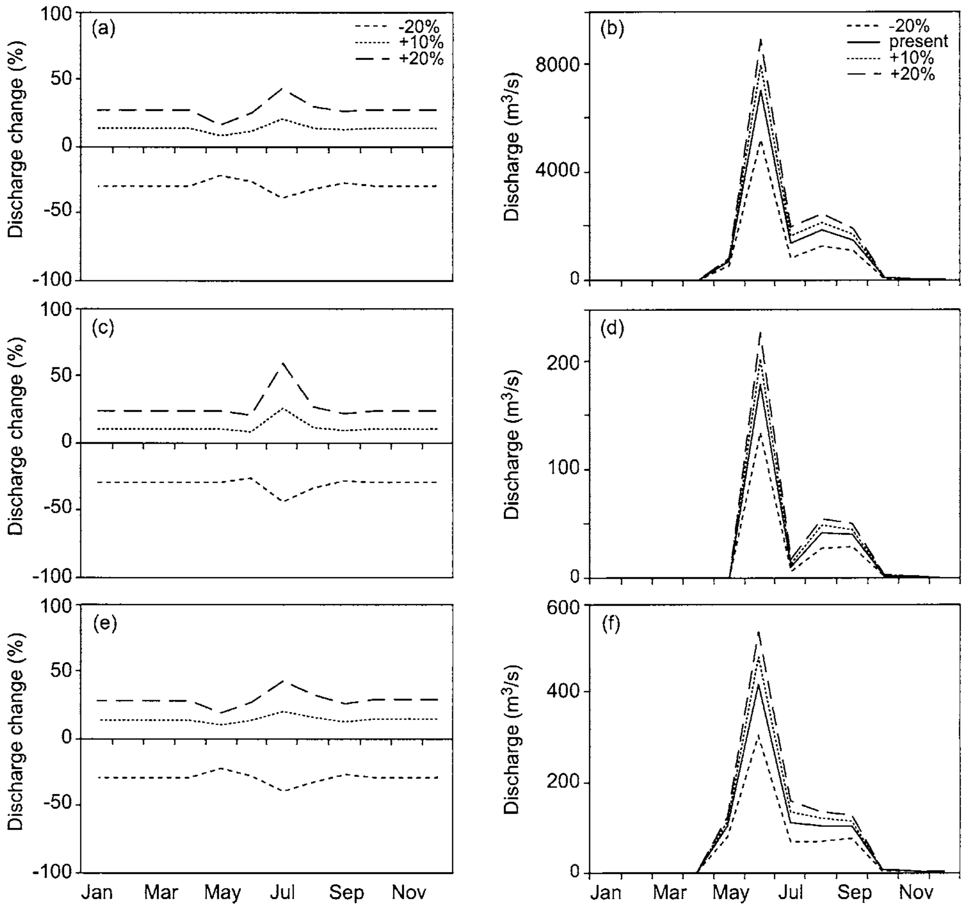

Figure 5. The effect of precipitation change on the change in discharge and on the absolute discharge for the Usa (a, b), the Khosedayu (c, d) and the Kosyu basin (e, f).

change in discharge is larger than the change in precipitation, both on annual and on monthly basis, because a smaller percentage of the available water evaporates than at present. The month of May is an exception, because the relative importance of evapotranspiration is low. In July, evapotranspiration is relatively important and the change in discharge is up to 50% higher than on annual basis. The discharge change in the Khosedayu is larger than for the other catchments, but the variation of the changes during the year are similar. The timing of discharge does not vary if the precipitation is changed.

In Figure 6 and Table II, the change in discharge with varying temperature is shown. On an annual basis, discharge increases when temperature decreases, because the period with water available for evapotranspiration becomes shorter. The opposite effect occurs for increased temperature. Discharge changes are up to 16%. In the Khosedayu catchment, the effect of reduced evapotranspiration is

Table II

Change in annual discharge compared to present-day discharge

	Change (%) $P-20\%$ $T-4\,°C$ $E-20\%$	Change (%) $P+10\%$ $T+2\,°C$ $E+10\%$	Change (%) $P+20\%$ $T+4\,°C$ $E+20\%$
Precipitation change			
Usa	–27	14	28
Khosedayu	–28	14	29
Kosyu	–28	14	29
Temperature change			
Usa	3	–4	–9
Khosedayu	16	–6	–15
Kosyu	3	–4	–7
Evapotranspiration change			
Usa	8	–4	–7
Khosedayu	8	–4	–7
Kosyu	9	–4	–8

largest. On a monthly basis, a change in temperature results in major discharge changes. When temperature is decreased by 4 °C, the runoff peak is delayed and the period with significant discharge lasts four months instead of five. In May (or June for the Khosedayu catchment) and September, discharge decreases by as much as 100%, because the groundwater storage is entirely depleted (May) and snowfall storage starts earlier than at present (September). Discharge increases in July by a factor 2.5 for the Usa and Kosyu and by a factor 14 for the Khosedayu due to a delay in the runoff peak. If temperature rises, the runoff peak becomes lower and is extended over May and June. In April, May and October, the relative increase in discharge is extremely high, because discharge at present is almost 0 m^3/s. Part of the snow has already melted in April and May, which causes the decrease in discharge in June and July.

Variations in evapotranspiration due to a change in vegetation have little effect on annual discharge (Table II). Discharge increases by about 8% if evaporation is 20% lower, and discharge decreases by 7% if evaporation is 20% higher. Also, the shape of the hydrograph is not affected by varying evapotranspiration. On a monthly basis, the change in discharge is similar to changes on annual basis, except in July (Figure 7). In this month, a decrease in evapotranspiration of 20% results in discharge which is about 20% higher than at present in Usa and Kosyu, while in Khosedayu the increase in discharge is 50%. The discharge change in this month

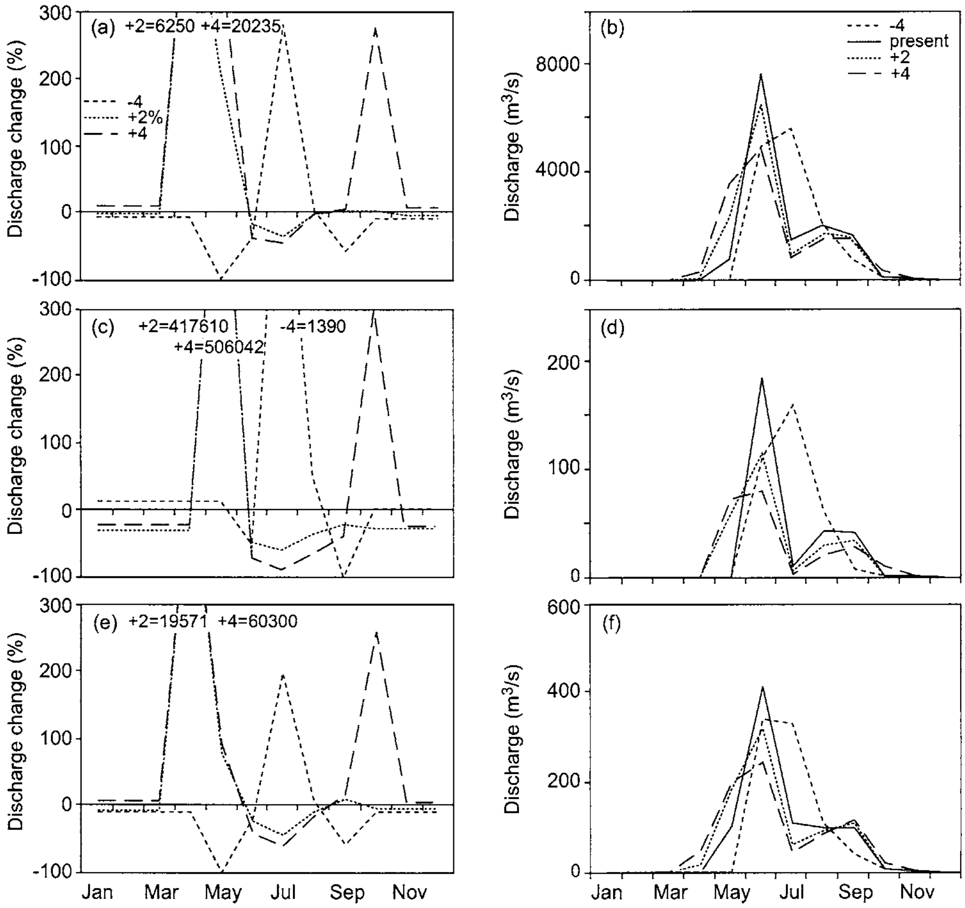

Figure 6. The effect of temperature change on the change in discharge and on the absolute discharge for the Usa (a, b), the Khosedayu (c, d) and the Kosyu basin (e, f).

is high, because evapotranspiration is highest and absolute discharge is low. An increase in evapotranspiration shows the opposite effect, although the decrease in discharge is smaller, i.e., a discharge decrease of 17% for Usa and Kosyu and of 25% for Khosedayu.

Varying the separation coefficient does not affect total annual discharge, but causes a minor change in the distribution within the year (Figure 8). A separation coefficient of 0 reduces discharge to zero from October to April. In the Khosedayu catchment, discharge decreases by 40% in July as well, while both other catchments show almost no effect. An increase in the separation coefficient by a factor 2 and 4 results in a higher discharge from October to April (100 and 300% higher). From May to September, the effect of a higher separation coefficient is small, except for Khosedayu where discharge increases are up to 130%.

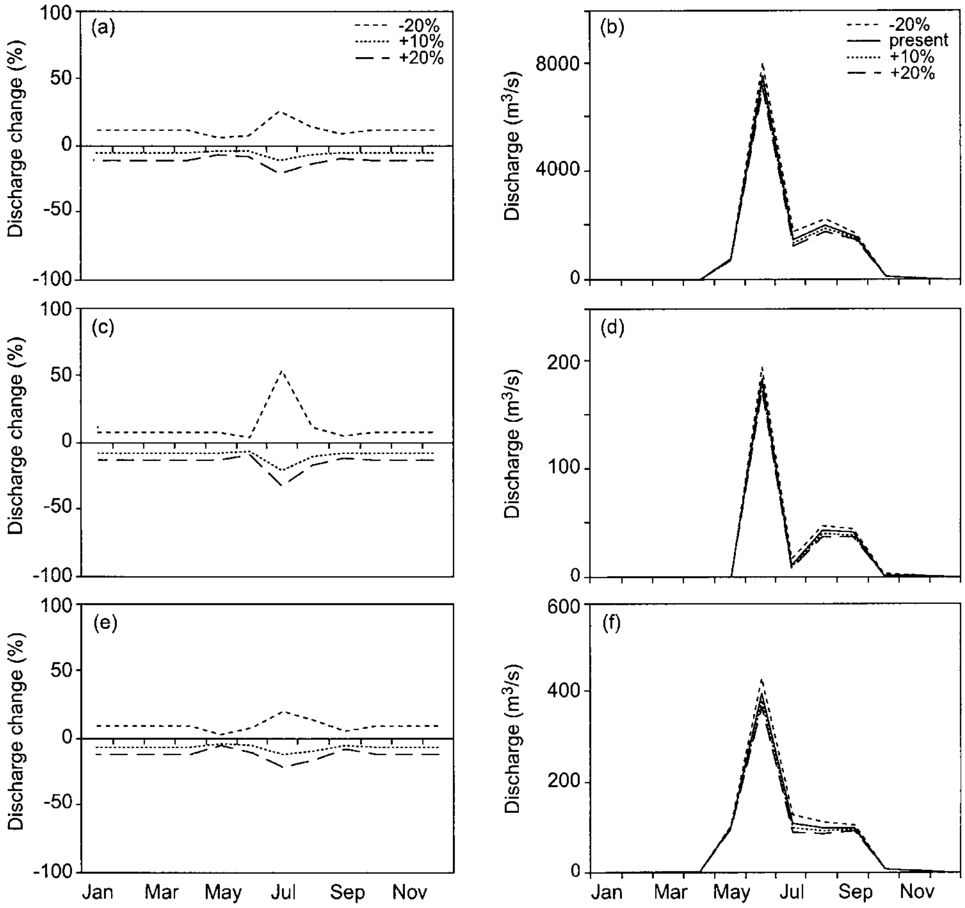

Figure 7. The effect of vegetation change on the change in discharge and on the absolute discharge for the Usa (a, b), the Khosedayu (c, d) and the Kosyu basin (e, f).

7. Results of the Scenario Study

Discharge simulated with the climate and environmental change scenario for the 2080s and the 2230s+ is shown in Figure 9. Annual discharge decreases by about 20% for the 2080 scenario (Table III). The spring runoff peak starts already in May, which is one month earlier than at present. The peak runoff (in June) is lower for all catchments. The volume of the snowmelt runoff, calculated over the months March until June decreases for all catchments by 16 to 22% (Table III). Discharge in summer (July until September) is lower for all catchments. Annual discharge simulated for the 2230 scenario increases by 10 to 16% compared to the present discharge. Peak discharge occurs in June and is lower than at present. The volume of snowmelt runoff is 12% larger in the Usa basin, 7% in the Khosedayu catchment and 19% in the Kosyu catchment compared to present-day values. In

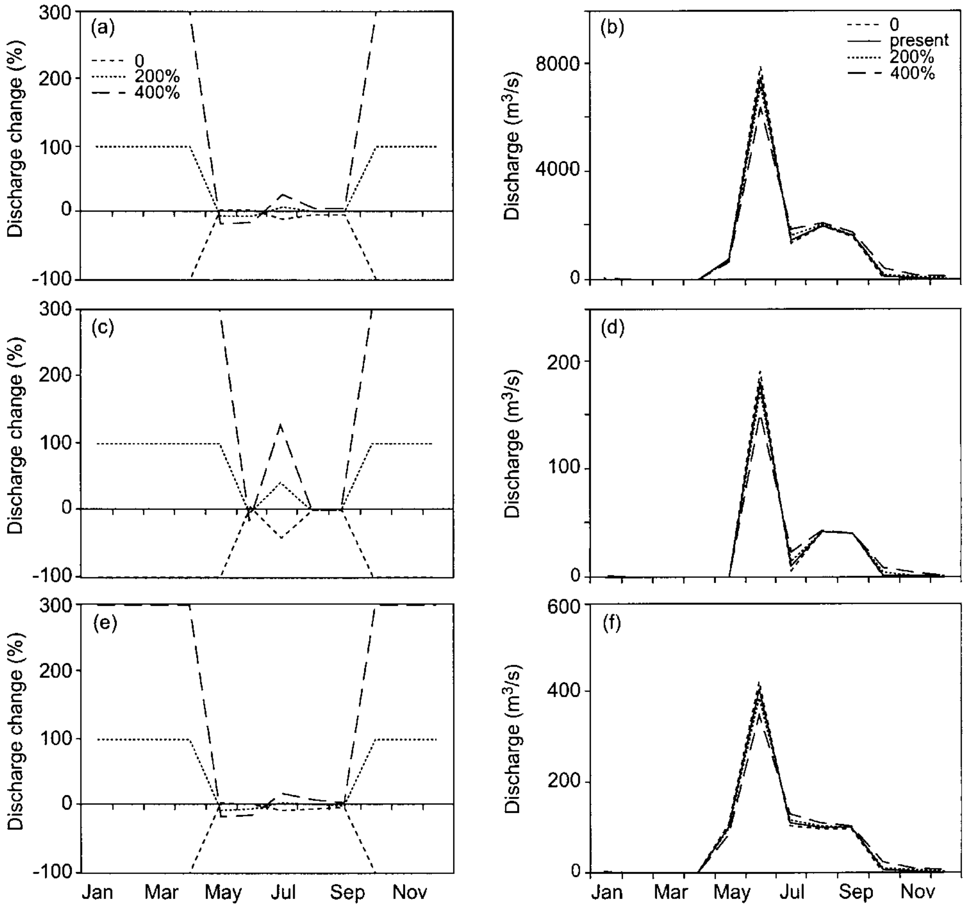

Figure 8. The effect of permafrost change on the change in discharge and on the absolute discharge for the Usa (a, b), the Khosedayu (c, d) and the Kosyu basin (e, f).

the Khosedayu catchment, summer discharge (July until September) is lower than at present and slightly higher than for the 2080 scenario, while discharge in both other catchments is almost similar to present-day discharge and higher than for the 2080 scenario.

8. Discussion

Discharge is highly sensitive to a change in precipitation. Especially the months May, June and July will be subject to discharge changes if precipitation changes. Discharge was found to increase more than proportionally with increasing precipitation in the Usa basin. Van Blarcum et al. (1995) found similar results for nine

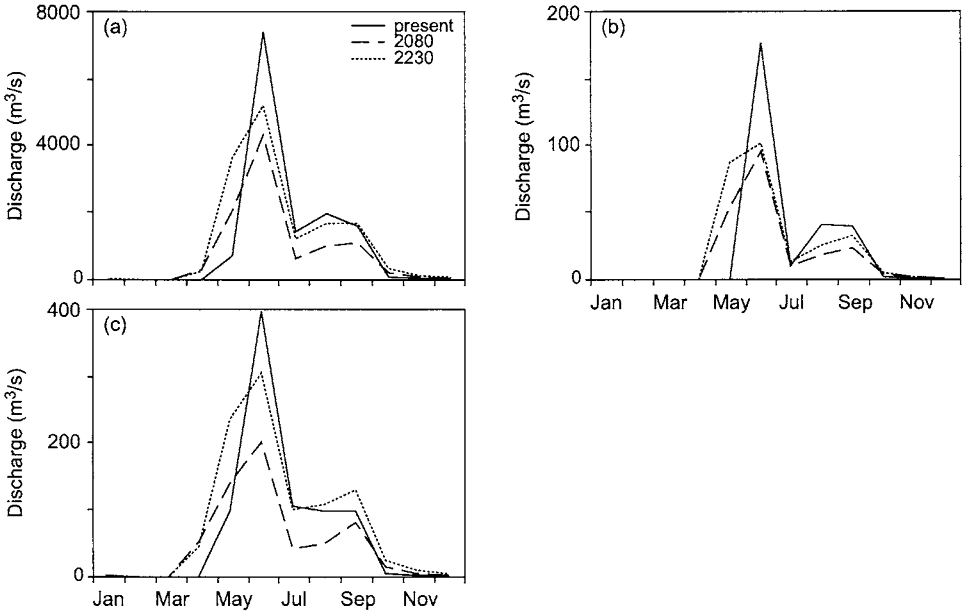

Figure 9. Present-day discharge and discharge simulated with the scenarios of 2080 and 2230 for the Usa (a), the Khosedayu (b) and the Kosyu basin (c).

Table III

Change in annual and peak discharge compared to present-day discharge with the climate change scenario

	Change for the 2080 scenario (%)	Change for the 2230 scenario (%)
Annual discharge		
Usa	–16	10
Khosedayu	–23	10
Kosyu	–20	16
Volume of snowmelt runoff		
Usa	–18	12
Khosedayu	–16	7
Kosyu	–22	19

river systems at high latitudes. Sensitivity to a precipitation change is almost equal in all catchments.

Discharge is also very sensitive to a change in temperature, which affects the timing and magnitude of the snowmelt runoff. In general, the effect of a temperature change on discharge is highest in periods where a temperature transition occurs from below to above the freezing point or vice versa. This effect is also found by Gleick (1987), Woo (1990) and Nijssen et al. (2001) in high latitude areas. From the results it appears that the coldest parts (e.g., the Khosedayu catchment) will experience the largest discharge changes, while discharge changes in relatively warmer areas (the Kosyu and Usa catchment) will be smaller.

The hydrological sensitivity to a change in vegetation and associated evapotranspiration is low. An increase in evapotranspiration leads to a smaller change in discharge than a decrease, because the soil moisture conditions become a limiting factor. A feedback mechanism may come into play when increased evapotranspiration decreases soil moisture contents to the extent that forest growth decreases. The sensitivity to a change in premafrost distribution is low. Nevertheless, discharge changes can be high during periods in which base flow is important, but absolute changes will be low, and the effect on the environment will be small.

The simulation with the scenario for the 2080s shows a decrease in annual discharge. The warmer climate shortens the season with snow accumulation and, as a consequence, runoff starts earlier in the season. The runoff peak decreases, as well as the total volume of snowmelt runoff. In summer, the increased evapotranspiration due to a temperature rise is not compensated by larger precipitation amounts, which causes discharge to decrease. The monthly hydrographs of the three catchments become smoother and discharge is distributed more evenly through the year. When precipitation amounts further increase as in the 2230 scenario, the higher evapotranspiration does not longer compensate the increase in precipitation, resulting in higher discharge. The higher winter precipitation results in a larger snow pack and consequently a larger volume of snowmelt runoff compared to present-day conditions. This is in agreement with Nijssen et al. (2001) who also found increased spring flow as a result of climate warming in cold snow dominated regions. In areas with high precipitation (Kosyu) the effect of a percentage change in precipitation will be highest. It is interesting to notice that the volume of snowmelt runoff decreases for the 2080s, but increases for the 2230s+ compared to present-day values. This suggests a non-linearity in response to climate forcing. This indicates that trends found in changing discharge in climate change impact studies may not be extrapolated over longer time periods.

The impact of climate change on river discharge in Arctic areas may be significant. Although Van Blarcum et al. (1995) and Nijssen et al. (2001) found an increase in discharge for high-latitude rivers, in this study simulation runs also showed discharge decreases. Increased winter flow due to degradation of permafrost and a shorter period with temperature below the freezing point may reduce ice formation on rivers (Woo, 1990). Also, the timing of break-up may be advanced.

The Khosedayu river is an example of a river where this process may occur as a result of future warming. A temperature increase may result in a shorter period with snow accumulation in winter, resulting in less snowmelt runoff. Although, if the increase in winter precipitation is sufficiently high, the volume of the snowmelt runoff will not change or become higher. These effects are enhanced in areas with high precipitation (for example, the Kosyu catchment). Evapotranspiration will increase due to a rise in summer temperature and the replacement of tundra vegetation with taiga forest, although the latter effect is small.

9. Conclusions

A modelling study to the sensitivity of the hydrological system in Arctic areas shows that these areas are sensitive to changes in climate and to a lesser extent to changes in related environmental variables (permafrost and vegetation). The effect of a change in precipitation on discharge is high throughout the year, while the influence of temperature is predominantly visible during the periods in which the temperature fluctuates around the freezing point (spring and autumn). The impact of a vegetation change on discharge is small. Sensitivity to a change in the distribution of permafrost is low on annual basis, but during periods of low flow changes in discharge can be high.

The HADCM2S750 scenario and related environmental changes projected to the 2080's show the following hydrological impacts. Enhanced evapotranspiration causes a decrease in annual discharge by about 20%. The total volume of snowmelt runoff and peak discharge decrease, because less snowfall occurs and the snowmelt runoff is extended over a longer period. Under equilibrium conditions, projected to the years after 2230, increases in evapotranspiration become smaller than the increases in precipitation, resulting in larger annual discharge. Due to the larger snow accumulation during winter, snowmelt runoff increases. These results suggest a non-linearity in basin response to climate change.

This study shows that the impact of climate change on the sensitive Arctic Usa basin can be substantial. Furthermore, the results suggest a non-linearity in basin response to climate change. Annual runoff, the amount of snowmelt runoff, and timing of ice break-up in the rivers may change, which is expected to have considerable implication for the environment and human society. This effect can be enhanced when climate variability increases, and the occurrence of extreme events becomes more important.

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