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Effects of changing climate on the hydrology of a boreal catchment and lake DOC — probabilistic assessment of a dynamic model chain

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Uncertainty in future climate projections is widely recognised, yet few impact model studies explore the implications of this uncertainty on catchment hydrological and biogeochemical responses. Here, we report a novel model chain using HBV, INCA-C and MyLake to simulate runoff, snow dynamics, ice cover, soil moisture, lake thermal stratification and in-lake DOC concentrations in Valkea-Kotinen, a headwater lake in a forest-covered boreal catchment. Impact response surfaces (IRSs) were constructed with 63 combinations of changes in temperature (–2 °C to +14 °C) and precipitation (–10% to +50%). Superimposing probabilistic projections of climate change onto the IRSs, we illustrate the uncertainty in impacts under projected climate change. Projected climate warming is likely to result in drier soils, shorter snow and ice periods, as well as earlier onset and longer duration of lake stratification. In contrast to earlier results, in-lake DOC concentrations are projected to decrease slightly (–6%), primarily due to decreased terrestrial runoff.

Introduction

Climate change is expected to have profound implications for boreal forests (Soja *et al.* 2007) and lakes (Keller 2007). Keller (2007) notes that a changing climate will lead to shorter periods of ice cover on boreal lakes, changed epilimnetic depths and longer periods of thermal stratification. Furthermore, the effects of climate change on the light and thermal regime of boreal lakes will be strongly influenced by changes in the input and cycling of dissolved organic carbon (DOC) (Keller 2007). Concentrations of DOC in surface waters are increasing in many parts of Fennoscandia and North America (Monteith *et al.* 2007, Vuorenmaa *et al.* 2006). The

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observed increase in DOC has been ascribed to reductions in acid deposition (Vuorenmaa *et al*. 2006, Monteith *et al*. 2007) and the effects of a changing climate (Lepistö *et al*. 2008, Haaland *et al*. 2010, Laudon *et al*. 2012). Increasing DOC concentrations have important effects on lake ecosystems. Higher DOC concentrations absorb light more effectively, changing the depth of the thermocline (Snucins and Gunn 2000). Shallower thermoclines and more effective thermal stratification can lead to increased hypolimnetic anoxia (Schindler *et al*. 1996). A combination of increased anoxia and slower rates of photodemethylation associated with greater UV absorbance by increased amounts of in-lake DOC may lead to higher methylmercury (MeHg) concentrations in biota (Hammerschmidt and Fitzgerald 2006). Oxygen-related changes induced by climate change may be more important than direct temperature changes for MeHg production (Verta *et al*. 2010) and accumulation in fish (Rask *et al*. 2010). Higher DOC concentrations are also linked to reductions in primary productivity (Arvola *et al*. 2014).

Climate has pronounced effects on lake DOC concentrations and the export of DOC from catchments. Increased precipitation can more effectively flush DOC from catchment soils and thereby increase lake DOC concentrations (Hongve *et al*. 2004, Haaland *et al*. 2010, Einola *et al*. 2011) while drought (Curtis and Schindler 1997) or warmer summers (Hudson *et al*. 2003) can lead to lower lake DOC concentrations as a result of longer water residence times and greater opportunity for in-lake consumption of DOC, which may increase rates at which MeHg is demethylated. Arvola *et al*. (2010) found an increase in water colour in response to a change in hydrological conditions after several dry summers.

A changing climate will also affect runoff. Higher temperatures in the boreal region will lead to less snow accumulation and earlier snowmelt, changing the annual distribution of runoff (Schindler *et al*. 1996). In Fennoscandia, the high flow period may switch from spring snowmelt to autumn and winter rain (Lawrence and Haddeland 2011). Higher temperatures will lead to a longer growing season, which may result in greater primary production, higher evapotranspiration and less runoff. These changes may have pronounced consequences for lake DOC budgets (Schindler *et al*. 1997, Keller 2007). While a range of future hydrological conditions have been projected, recent research suggests drier conditions and lower lake levels in parts of Fennoscandia (Wetterhall *et al*. 2011).

While the uncertainty of future climate conditions and corresponding catchment responses is acknowledged by all, explicit treatment of this uncertainty is rare: most published model studies often include only a handful of scenarios to explore future uncertainties (e.g. Posch *et al*. 2008, Futter *et al*. 2009, Saloranta *et al*. 2009, Oni *et al*. 2012). Increasing numbers of climate model simulations being conducted have allowed attempts to express regional climate uncertainty probabilistically (e.g. Räisänen and Ruokolainen 2006, Prudhomme *et al*. 2010, Frieler *et al*. 2012). This study uses one such probabilistic projection of regional climate change (Harris *et al*. 2010) and translates it to the response of a chain of three impact models that simulate the DOC concentration of a lake in southern Finland. In this work, we follow the approach of authors who have utilized the results of the ENSEMBLES project to allow for a more explicit treatment of climate change uncertainty (Fronzek *et al*. 2011, Weiss 2011, Wetterhall *et al*. 2011). The purpose of this study was to demonstrate the impact response surface approach to quantify the range of uncertainty introduced by climate uncertainty into impact modelling. We did not attempt to provide a full assessment of uncertainty propagated through the entire model chain. To our knowledge, this study is the first report of a probabilistic approach to a model chain applied to examine potential impacts of climate change on the hydrology of a boreal headwater catchment and its effects on terrestrial DOC export and in-lake DOC concentrations.

Material and methods

Site

Valkea-Kotinen is a small headwater catchment of the Kokemäki river water system. The catchment is located in southern Finland (61°14´N,

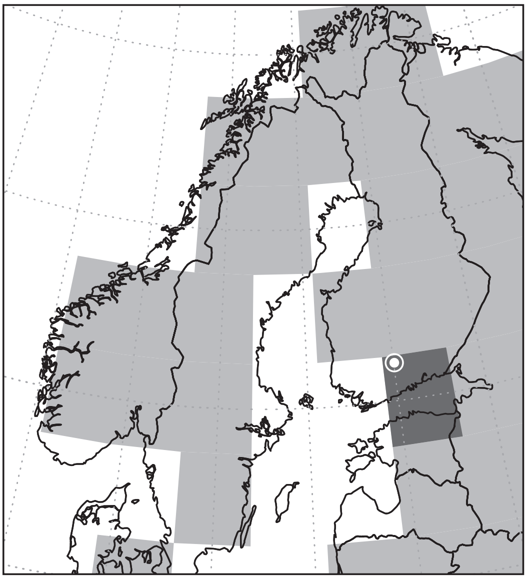

Fig. 1. Location of the catchment Valkea-Kotinen in Lammi, southern Finland (white circle), for which impact model chain simulations were performed and the climate model land grid box (dark grey) for which probabilistic climate projections were obtained (other land grid boxes in light grey).

25°04´E; Fig. 1) in a protected area of old-growth forest and receives only background levels of long-range transported air pollution. The catchment and monitoring activities are described in Bergström *et al*. (1995), Ukonmaanaho *et al*. (1998) and Vuorenmaa *et al*. (2014). Stream flow is monitored by the Finnish Environment Institute at the catchment outflow and water chemistry monitored in the lake and at the catchment outflow.

Valkea-Kotinen is typical of glaciated boreal landscapes. The catchment (30 ha) contains areas of forested mineral soil, forested and open peatlands and a discharge lake (156 m a.s.l) with stream. The lake subcatchment is 22 ha and the lake area 4 ha. The mineral soils in the catchments are predominately Podzols, developed on shallow glacial drift (till) deposits (Starr and Ukonmaanaho 2001). The forest cover consists mainly of old-growth mixed stands of Norway spruce, birch and aspen.

The Valkea-Kotinen catchment has been the subject of much biogeochemical research (Keskitalo *et al*. 1998, Starr and Ukonmaanaho 2001, 2004, Arvola *et al*. 2010, Einola *et al*. 2011, Vuorenmaa *et al*. 2014). Detailed data on soil and stand carbon stocks, litter production and soil-water DOC concentrations are available (*see* Starr and Ukonmaanaho 2001, 2004). A number of limnological studies focussed on in-lake carbon cycling and greenhouse gas production have also been conducted (Vähätalo *et al*. 2003).

Meteorological data used in this study were obtained from nearby Finnish Meteorological Institute weather stations; average daily temperature (T) and daily precipitation (P) data from Lammi Pappila (61°03´N, 25°03´E, 125 m a.s.l.), and P, T and total global radiation data from Jokioinen (60°49´N, 23°30´E, 104 m a.s.l.). The Lammi station is closer to the Valkea-Kotinen site but the data from Jokioinen were also used as radiation data were not available at Lammi. For 1980–2012, the long-term mean annual temperature at Lammi Pappila is +4.2 °C and the annual amount of precipitation is 645 mm (Pirinen *et al*. 2012). Approximately 30% falls as snow or sleet, giving rise to a distinct snowmelt peak in runoff at the end of April. Global radiation averages 3315 MJ m^{-2} yr^{-1} at Jokioinen for 1981–2010 (Pirinen *et al*. 2012).

Precipitation in the region has increased slightly during the past century but the increase is not statistically significant (Ylhäisi *et al*. 2010). The Lammi meteorological station has reported a statistically significant warming of almost 0.4 °C per decade (1964–2011). For the shorter period 1990–2010 the linear trend is weaker and covered by noise. The lake Valkea-Kotinen is covered by ice for about 170 days on average (1990–2011), and the ice period shortened by about 15 days per decade for the same period (Jylhä *et al*. 2014).

Warming is expected to continue in the region throughout the 21st century by 0.3–0.4 °C per decade, or 0.4–0.5 °C per decade in case of scenario A2, with the warming projected to be strongest in winter. The warming may reach 0.4–0.5 °C per decade according to scenario A2. Precipitation is expected to increase relatively more in winter than in summer, while summer precipitation will remain more abundant than winter precipitation. The proportion of precipitation as snow will decrease and snow cover is expected to become less common, especially in early and late winter (Jylhä *et al*. 2014).

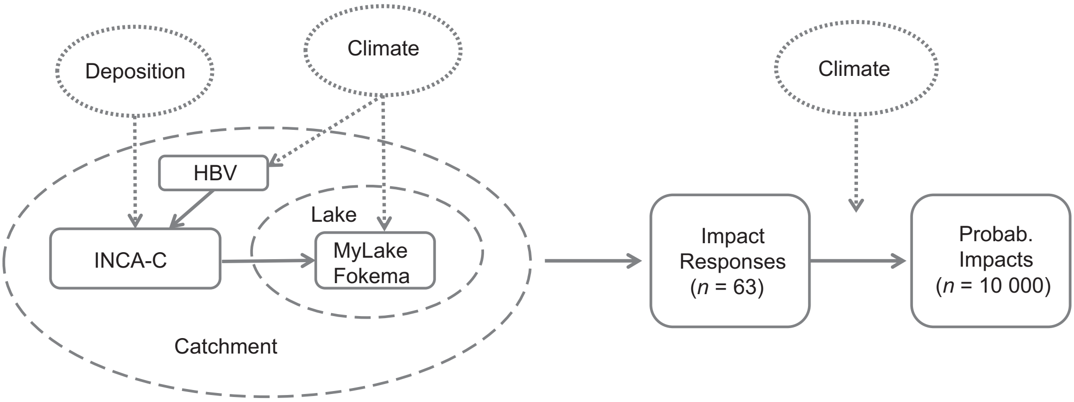

Fig. 2. A diagram of impact model chain applied to the Valkea-Kotinen catchment.

A decrease in acidifying deposition has been recorded since 1988. In 1988–1997, the mean depositions of SO_4^{2-} and NO_3^- were 419 and 205 mg m^{-2}, respectively. In 2003–2011, the mean deposition levels decreased to about 170 mg S m^{-2} and about 160 mg N m^{-2} (Ruoho-Airola *et al*. 2014).

Impact model chain

A model chain consisting of the hydrological model HBV (Bergström 1992, Sælthun 1996), INCA-C that simulates the carbon cycle at the catchment-scale (Futter *et al*. 2007), the lake model MyLake (Saloranta and Andersen 2007) extended with the FOKEMA submodule that simulates the biological decay and photomineralization of DOC in surface waters was used to estimate the possible effects of future climate change on lake ecosystem function (Fig. 2).

HBV

The HBV model (Bergström 1976, Bergström 1992, Sælthun 1996) is a catchment-scale, daily time step rainfall-runoff model for simulating water balances. The model version used here requires inputs of daily time series of precipitation and air temperature (Sælthun 1996). It produces daily time series of runoff which can be calibrated against measured values. Depending on air temperature, precipitation in HBV can either enter the soil or be accumulated as snow. The model has a degree day snowmelt routine. Precipitation interception is land-cover dependent. Precipitation entering the soil can contribute either to runoff or recharge. The runoff hydrograph is dependent on three time constants representing quick, soil water and groundwater flows. Potential evapotranspiration (PET) in HBV is simulated using a degree day model. The actual evapotranspiration (AET) is estimated as a function of PET and soil moisture, where rates of AET are slower in drier soils (Sælthun 1996). HBV produces outputs of soil moisture deficits (SMD), runoff, snow depth expressed as snow water equivalents (SWE) and hydrologically effective rainfall (HER). The latter term is an estimate of the depth of water entering the soil that eventually contributes to runoff. It is equivalent to rainfall plus snowmelt minus interception and actual evapotranspiration. The SMD term is equal to the difference between current estimated water content and the user-specified field capacity (FC). The duration of snow cover was estimated as the number of days in a year in which HBV simulated a non-zero snow depth. Despite its relative simplicity, HBV has been widely applied for simulating runoff in Fennoscandia (Bergström 1992, Sælthun 1996, Lawrence and Haddeland 2011). The model has also been used in numerous climate change studies (Futter *et al*. 2009, Lawrence and Haddeland 2011, Wetterhall *et al*. 2011, Oni *et al*. 2012). HBV was calibrated to measured stream flow in the catchment outlet from 1990–2007, with the strategy to maximize the Nash-Sutcliffe (NS) coefficient (Nash and Sutcliffe 1970).

INCA-C

INCA-C is a conceptual model of carbon cycling in soils and surface waters (Futter *et al.* 2007). The model operates at a daily time step. INCA-C has been applied to headwater boreal forest catchments in Canada (Futter *et al.* 2007), Finland (Futter *et al.* 2008, 2009), Sweden (Futter *et al.* 2011) and Norway (Futter and de Wit 2008). The model simulates key processes in the forest carbon cycling including litterfall, incorporation of litter into soil organic carbon, leaching and mineralization of soil organic carbon as well as microbial and photolytic decomposition of organic carbon in surface waters. All in-soil process rates are dependent on soil moisture and temperature. Soil moisture is obtained from the HBV-estimated SMD while soil temperature is estimated using the model of Rankinen *et al.* (2004). Organic carbon solubility in the soil is dependent on SO_4^{2-} concentrations (Futter *et al.* 2009). The INCA-C processes and equations are documented in Futter *et al.* (2007, 2009).

The INCA-C model has been used to simulate possible effects of future climate on DOC in headwater streams in Canada (Aherne *et al.* 2008) and Finland (Futter *et al.* 2009) as well as large rivers in Canada (Oni *et al.* 2012).

The model calibration presented here is based on Futter *et al.* (2009). INCA-C was calibrated against measurement data from 1990–2007: DOC in the lake and catchment outflow as well as to stream flow at the catchment outflow. EMEP SO_4^{2-} deposition was used as a surrogate for soil solution SO_4^{2-} concentrations. Deposition of SO_4^{2-} equal to current legislated emissions (CLE) was used when running INCA-C to generate climate scenarios. The deposition scenarios are described in Posch *et al.* (2008). Soil carbon inputs in INCA-C are assumed to be derived primarily from litterfall. Total canopy litterfall was estimated using empirical relationships between climate and stand conditions for Norway spruce (Starr *et al.* 2005) and Scots pine (Saarsalmi *et al.* 2007). Average stand properties were obtained from table 3.9.1.1 of Bergström *et al.* (1995). It was assumed that the old-growth forest is at steady state. A Monte Carlo analysis based on Latin Hypercube sampling of the parameter space (Futter *et al.* 2007) was performed to assess uncertainty in INCA-C predictions. The sampled parameter range was based on the best-performing manual calibration ±20%. The parameter space was sampled 10 000 times and the 20 parameter sets with highest joint NS statistics for lake and catchment outlet DOC were retained for further analysis.

MyLake

MyLake is a one-dimensional, daily time step lake model (Saloranta and Andersen 2007, Saloranta *et al.* 2009). The primary function of MyLake is to simulate lake thermal processes which are forced by time series of air T, P, solar radiation, wind speed and relative humidity. MyLake simulates lake stratification and seasonal lake ice cover and it is well suited for multi-decadal simulations to evaluate the possible impacts of climate change on lake ecosystem functioning (Saloranta and Andersen 2007). It has been used previously in climate change impact studies (e.g. Forsius *et al.* 2010, Dibike *et al.* 2011). MyLake only simulates the most significant physical, chemical and biological processes controlling lake response to climate. As described in Saloranta and Andersen (2007), the lake dynamics follows equations presented by Ford and Stefan (1980), Riley and Stefan (1988), and Hondzo and Stefan (1993), while the temporal development of ice is based on Leppäranta (1991) and Saloranta (2000). Stability as defined by Schmidt is calculated following equations presented by Korhonen (2002). Saloranta and Andersen (2007) report a sensitivity and uncertainty analysis of the application of MyLake to phytoplankton and inorganic phosphorus in Vansjø-Storefjorden, Norway. They found catchment loading of phosphorus to be the most important forcing factor to focus on to improve model simulations (Saloranta and Andersen 2007). In a Markov chain Monte Carlo uncertainty analysis, Saloranta and co-workers conclude that the uncertainties connected to the MyLake model parameter values are smaller than those connected with the climate forcing (Saloranta *et al.* 2009). Dibike and co-workers applied MyLake with default parameter values to broad regional patterns of changes in lake-ice conditions and concluded it was appropriate for modelling future changes (Dibike *et al.* 2011).

In the version of MyLake used here, DOC is input from the surrounding catchment and can be consumed in the lake through both photochemical and biological processes. The bacterial decay of DOC is simulated in a submodule (FOKEMA) which calculates microbial mineralization following a three-pool kinetic equation as described in Vähätalo *et al.* (2010). Incoming DOC is divided into three pools with different propensity for bacterial mineralization. The first pool has the fastest decay rate, the second pool a lower rate and the third pool, which constitutes the largest fraction of DOC, is not affected by bacterial mineralization. MyLake was calibrated against DOC concentrations measured in the uppermost layer (0–1 m) of the lake during 1990–2005. The optimal parameter values corresponding to the best fit were chosen so as to maximize both correlation between daily measured and modelled DOC, and the NS coefficient. The model sensitivity was investigated by calculating the range in modelled lake DOC obtained by varying model parameters and incoming DOC flux. The parameter values were sampled at random from a normal distribution with mean corresponding to optimal parameter values and standard deviation of one third of the optimal values. The incoming DOC flux was varied by factors of 0.1, 0.5, 1.5 and 3.

The onset of lake stratification is here defined at a stability threshold of 1 J m^{-2}. The stratified period is counted from the day when the stability exceeds 1 J m^{-2} for at least 30 consecutive days. The stratified period ends when the stability is lower than 1 J m^{-2} for the first time.

Probabilistic impacts response analysis

A probabilistic analysis of future impacts was conducted by first constructing impact response surfaces (IRSs) that depict the sensitivity of the impact model chain to changes in T and P and combining these with probabilistic projections of climate change for future time periods (Fronzek *et al.* 2010). The sensitivity of the impact model chain to climate change was determined by changing input T and P by a set of perturbations (ΔT, ΔP) spanning a range of possible future climates. ΔT was varied between –2 °C and 14 °C at 2 °C intervals; and ΔP between –10% and 50% at 10% intervals, resulting in a grid of 63 unique ΔT, ΔP combinations. Seasonal scalers for temperature and precipitation were applied to these perturbations that define stronger changes during winter (Table 1). The scalers were calculated as the mean seasonal change divided by the mean annual change for all 20-year periods throughout the 21st century in the Harris *et al.* (2010) data set. The sensitivities of the impact responses to climate change were plotted as IRSs with contour lines using the contour function in MATLAB R2012b with ΔT, ΔP coordinates.

Joint probabilities of long-term changes in annual mean T and P relative to the baseline period 1961–1990 for the periods 2020–2039, 2050–2069 and 2080–2099 were obtained from Harris *et al.* (2010). These T and P estimates combine information from a perturbed physics ensemble with multi-model ensembles of general circulation models. The probabilistic climate projections quantify the uncertainties in leading physical, chemical and biological feedbacks. The data were provided as joint frequency distributions with a sample size of 10 000 for the "Gulf of Finland" grid cell in which the Valkea-Kotinen catchment is situated (Fig. 1) (Harris *et al.* 2010).

The IRSs were evaluated at the locations of the 10 000 pairs of ΔT, ΔP of future time periods by linearly interpolating from the 63 impact model results. For each ΔT, ΔP combination in the climate projections, the impact response was calculated by averaging the impact model results for the four nearest nodes of the delta change

Table 1. Seasonal scalers for temperature and precipitation changes calculated from all 20-year periods between 2000 and 2099 of the Harris *et al.* (2010) dataset for the "Gulf of Finland".

Season	Offset	
	Temperature	Precipitation
Winter (DJF)	1.276	1.290
Spring (MAM)	0.947	1.126
Summer (JJA)	0.844	0.825
Autumn (SON)	0.933	0.760

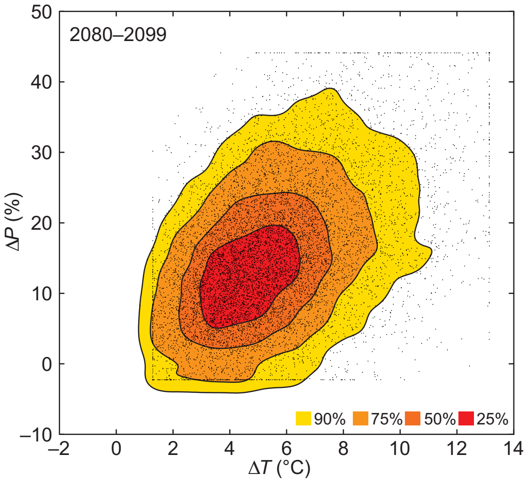

Fig. 3. Changes from 1961–1990 to 2080–2099 in annual mean temperature (°C) and precipitation (%) for the Grand Ensemble probabilistic projection for the land grid box Gulf of Finland (Harris *et al.* 2010). The scattered dots depict the 10 000 combinations of change in temperature and precipitation (ΔT, ΔP). The contours of the coloured areas represent different densities of the scatter. Probabilities are represented as the percentage of projections enclosed within closed contour lines; 90% of the projections are within the outermost contour line, 75% and 50% within the intermediate, and 25% within the innermost contour line.

combinations, using the inverse distances to the nearest nodes as weighting coefficients.

The probabilities of the impact responses were estimated as cumulative frequencies, which were determined for the 10 000 linearly interpolated values for each time period. The IPCC guidance note on treatment of uncertainty recommends a likelihood scale in which likely and very likely correspond to 66%–100% and 90%–100% probability, and unlikely and very unlikely to 0%–33% and 0%–10% probability (Mastrandrea *et al*. 2011). The 10th, 33rd, 50th, 66th, and 90th percentiles of the cumulative frequency distributions were calculated in MS Excel 2010 for the impact response variables length of catchment snow period, length of lake ice period, onset of thermal stratification and length of stratified period, soil moisture deficit, annual runoff to lake, DOC concentrations in influx to lake, DOC flux to lake, and DOC concentrations in the top 0–1 m of lake water.

To illustrate the joint probability distribution of projected future climate change, the probability densities of the changes from 1961–1990 to 2080–2099 are shown as a scatter plot of individual (ΔT, ΔP) values combined with four contour lines that represent certain percentages of the points within the lines (Fig. 3). For each time period, the distribution of the 10 000 joined (ΔT, ΔP) values was estimated with a bivariate kernel density estimator (Botev *et al*. 2010) which produced density values for each (ΔT, ΔP) combination. Contours were calculated by comparing the relative sum of densities in relation to the total density sum to certain fixed percentages (25th, 50th, 75th, and 90th percentiles). The probabilities are visualized as coloured areas between contour lines that represent the probability expressed as percentage of the total sample (Fig. 3). The filled contours were plotted with the 'contourf' function in MATLAB R2012b. If the frequency of T perturbations are calculated separately (without regard for P perturbations), the frequency analysis shows that a warming of less than 4 °C is unlikely (< 33% of the 10 000 projections of $\Delta T < 4$ °C) and a warming of at most 6 °C is likely (> 66% of the 10 000 projections of $\Delta T < 6$ °C). Similarly, for individual frequencies of P perturbations, the corresponding observations are that it is unlikely that P will increase by less than 11% and likely that P will increase by most 19%. As compared with the joint probability graph (Fig. 3), the individual frequency analysis gives a narrower view of future change which shows the advantage of using joint probability distributions providing more comprehensive information in the background of impact response analysis.

Results

Model calibrations and parameter sensitivity

The HBV calibration from Futter *et al*. (2009) against measurement data from the catchment outlet was used here. The calibration was able to achieve a NS statistic of 0.65. This is comparable to model goodness-of-fit for other HBV applications to small boreal headwater catchments (Futter *et al*. 2007, 2008, 2011). The INCA-C calibration from Futter *et al*. (2009) was re-visited in an attempt to improve the model fit to observed DOC concentrations in the lake. The final calibration achieved a NS statistic of 0.443 and 0.293 for DOC in the lake and catchment outflow respectively. The range in predicted DOC concentrations from the Monte Carlo analysis, expressed as (maximum – minimum)/(average predicted value for each day) × 100, was between 0% and 40%, with the average values of 7.5% and 15% for the lake and catchment outflow, respectively.

In the calibration of the bacterial decay parameters in FOKEMA, one fifth of the samples (42 of 227) were acceptable, with correlation coefficients > 0.4 and NS coefficients > 0.01. For the best fit, the correlation between measured and modelled daily values was 0.48 and the NS coefficient was 0.08. The model was more sensitive to incoming DOC than to the decay parameter values. Of the decay parameters, the fastest decay rate was the most important. For the accepted model runs, the range (maximum–minimum) of simulated DOC corresponded to 7% of the predicted average value.

Climate sensitivity of impact models and probabilistic analysis

Catchment snow cover, lake ice cover and lake stratification

The length of the HBV-estimated snow period in the catchment depends almost only on changes in temperature. The frequency analysis of the model chain results indicate that it is likely that the length of the snow period will decrease by 74% from 179 to 46 days (Table 2). A shorter snow period than 23 days is unlikely. The length of the ice period is likely to decrease towards the end of the century by 81% (from 166 to 32 days). It is unlikely that the ice period will be shorter than 12 days. The stratification of the lake water will begin earlier, likely by 23 April, as compared with 12 May in reference conditions, and the length of the stratified period is likely to increase, from 131 days in reference conditions to 174 days. It is unlikely that the onset of stratification will occur earlier than 17 April, or that the length of the stratified period will be shorter than 162 days (Table 2).

Seasonal patterns in actual evapotranspiration and runoff

There is a clear seasonal pattern in the hydrologic response to climate change. For each season separately [winter: Dec–Jan–Feb (DJF), spring: Mar–Apr–May (MAM), summer: Jun–Jul–Aug (JJA), and autumn: Sep–Oct–Nov (SON)], the standardized values of AET and runoff were calculated as the ratio of the seasonal average divided by the value for reference conditions (no climate change ΔT, $\Delta P = 0, 0$). The seasonal responses of AET (Fig. 4) and runoff (Fig. 5) were plotted as contours against annual changes

Table 2. Summary of model chain results for 2080–2099 compared with the reference values, 15-year median values for the period 1990–2005. Change is expressed as percentiles of 10 000 interpolated values, based on modelled results for 63 combinations of ΔT, ΔP and 10 000 probabilistic projections of ΔT, ΔP.

Modelled response	Unit	Reference	Percentiles of change by 2080–2099 compared with reference				
			10th	33rd	50th	66th	90th
Length of snow period	days	179	–94%	–87%	–77%	–74%	–44%
Length of ice period	days	166	–94%	–93%	–84%	–81%	–32%
Onset of stratification		12 May	8 April	17 April	22 April	23 April	30 April
Length of stratified period	days	131	18%	24%	24%	33%	55%
Soil moisture deficit	mm	34	3%	12%	19%	25%	51%
Annual runoff to lake	mm	232	–23%	–10%	–2%	5%	27%
DOC conc. in influx to lake	mg l^{-1}	35.6	–3%	0%	2%	4%	9%
DOC flux to lake	g m^{-2} yr^{-1}	10	–13%	–6%	–2%	2%	14%
DOC conc. in lake 0–1 m	mg l^{-1}	14.3	–10%	–7%	–6%	–5%	–2%

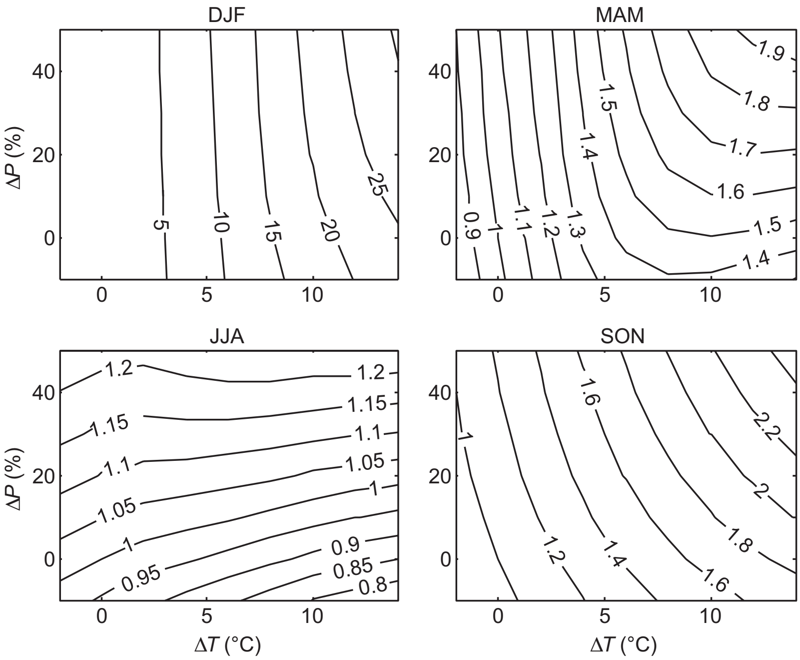

Fig. 4. Seasonal evapotranspiration. HBV-modelled average actual evapotranspiration (AET) in winter (DJF), spring (MAM), summer (JJA) and autumn (SON) in response to changes in annual temperature and precipitation. Results are given as standardized values: a value of one represents AET in reference conditions (DJF 0.03 mm d^{-1}; MAM 0.51 mm d^{-1}; JJA 1.04 mm d^{-1}; SON 0.31 mm d^{-1}).

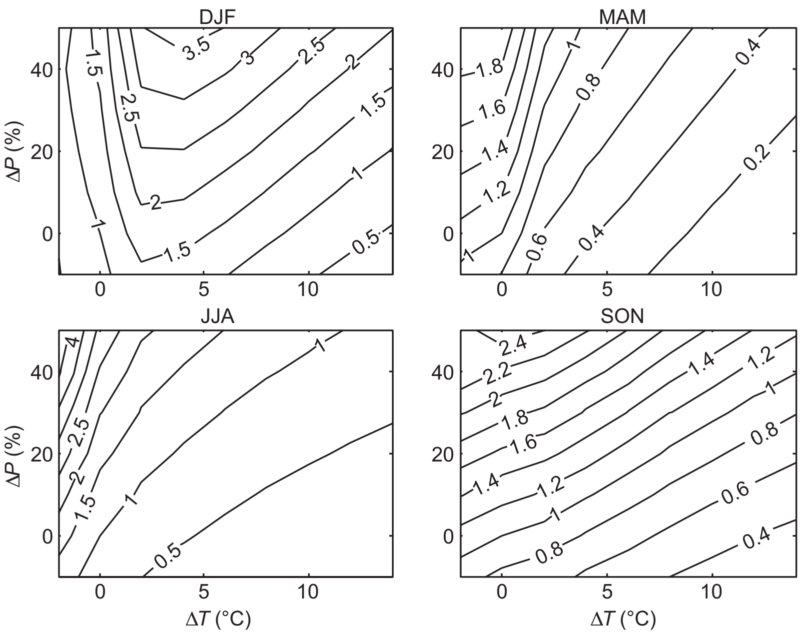

Fig. 5. Seasonal runoff. HBV-modelled average runoff in winter (DJF), spring (MAM), summer (JJA) and autumn (SON) in response to changes in annual temperature and precipitation. Results are given as standardized values: a value of one represents runoff in reference conditions (DJF 0.42 mm d^{-1}; MAM 1.08 mm d^{-1}; JJA 0.32 mm d^{-1}; SON 0.36 mm d^{-1}).

in temperature (ΔT, x-axis) and precipitation (ΔP, y-axis). Vertical contour lines indicate that the pattern is controlled purely by temperature, whereas horizontal contour lines are indicative of pure precipitation control. In winter (DJF), the AET response pattern is controlled primarily by the temperature scenario, whereas the reverse is true for the summer (Fig. 4 JJA). Wintertime changes in AET are manifold as compared with those in summer. In spring, up to a certain point, the AET change is controlled by increasing temperature, but above about a 5 °C change in T, precipitation is the major control (Fig. 4: MAM). Changing runoff is relatively more dependent on changing precipitation (Fig. 5). In winter, runoff change is controlled primarily by temperature up to ΔT about 3 °C, after which the regulating role of ΔP increases (Fig. 5 DJF).

Soil moisture and annual runoff

A visual impression of the probability of future HBV-estimated soil moisture deficit (Fig. 6a) is obtained by comparing three panels showing the IRSs combined with climate probabilities for three time slices (2020–2039, 2050–2069, 2080–2099) (Fig. 6a). The IRSs, plotted as labelled contour lines, show the sensitivity of the modelled soil moisture deficit to changes in climate (ΔT, ΔP). The sensitivity contour lines are the same for all time slices, as the sensitivity does not change with time. Soil moisture deficit is sensitive to both temperature and precipitation change, which can be seen in that the IRS contour lines are not parallel to either of the ΔT or ΔP axes. An analysis of monthly changes (not shown) showed the largest increases in SMD in April and May as a result of warmer simulated temperatures and an earlier onset of the simulated growing season.

The probability distribution of future climate change varies with time, however. The coloured filled climate contours in the three panels that represent the time slices 2020–2039, 2050–2069 and 2080–2099 (Fig. 6) shift with time towards higher warming and larger increase in precipitation. In addition, as time evolves, the uncertainty in the climate projections grows larger — the area of the filled contours expand. For the time period 2080–2099, the climate contours are identical to the coloured filled contours plotted in the same graph as the scatter plot of the joint probability distribution ΔT, ΔP (Fig. 3).

At the same time as the climate contours shift from left to right, the probability of drier soil (larger SMD) increases. This can be seen in the bottom panel (2080–2099), where the innermost climate contour is far enough to the right to intersect the SMD contour line of 1.2. The implication is that towards the end of the century there is a higher probability for soils to be drier (by a factor of 1.2) as compared with the soils in reference conditions (SMD 34 mm; Table 2).

Annual HBV-estimated runoff (Fig. 6b) is sensitive to changes both in temperature and precipitation and the IRS contour lines are evenly distributed over the ΔT, ΔP space. The slope of the annual runoff IRS is less steep than for the seasonal runoff (Fig. 5). The changes in the aspect of the slope that are striking in the seasonal case, especially for winter runoff (Fig. 5; DJF) are missing at the annual level. The contour line of one, representing annual runoff in reference conditions (232 mm; Table 2), cuts the probabilistic climate contour approximately in half, which means that in this analysis, the probability for an increase in annual runoff is no larger than that of a decrease.

DOC concentrations

INCA-C estimated DOC concentrations in the influx to the lake (Fig. 6c) and MyLake/FOKEMA estimated concentrations of DOC in the lake (Fig. 6d) vary with both ΔT and ΔP. The DOC IRS plots have less steep slopes than the SMD or runoff IRSs, indicating lower sensitivity to climate change. The influx DOC concentration in reference conditions is 35.6 mg l^{-1}. The corresponding contour line lies near the upper border of the innermost climate contour in the lowest panel (2080–2099), which would mean a slight increase in DOC influx to the lake (Fig. 6c). For the lake DOC concentrations (Fig. 6d), future climate conditions may lead to slightly lower values than in reference conditions (14.3 mg l^{-1}; Table 2). A frequency histogram with cumulative frequency distribution (Fig. 7) shows the shift of in-lake DOC concentrations towards lower values.

Frequency distributions of model chain results

Reference values for the impact response variables are presented together with the 10th, 33rd, 50th, 66th, and 90th percentiles of their frequency distributions calculated for the time period 2080–2099 (Table 2). Using the likelihood scale recommended by IPCC, our results indicate that it is likely that DOC concentrations

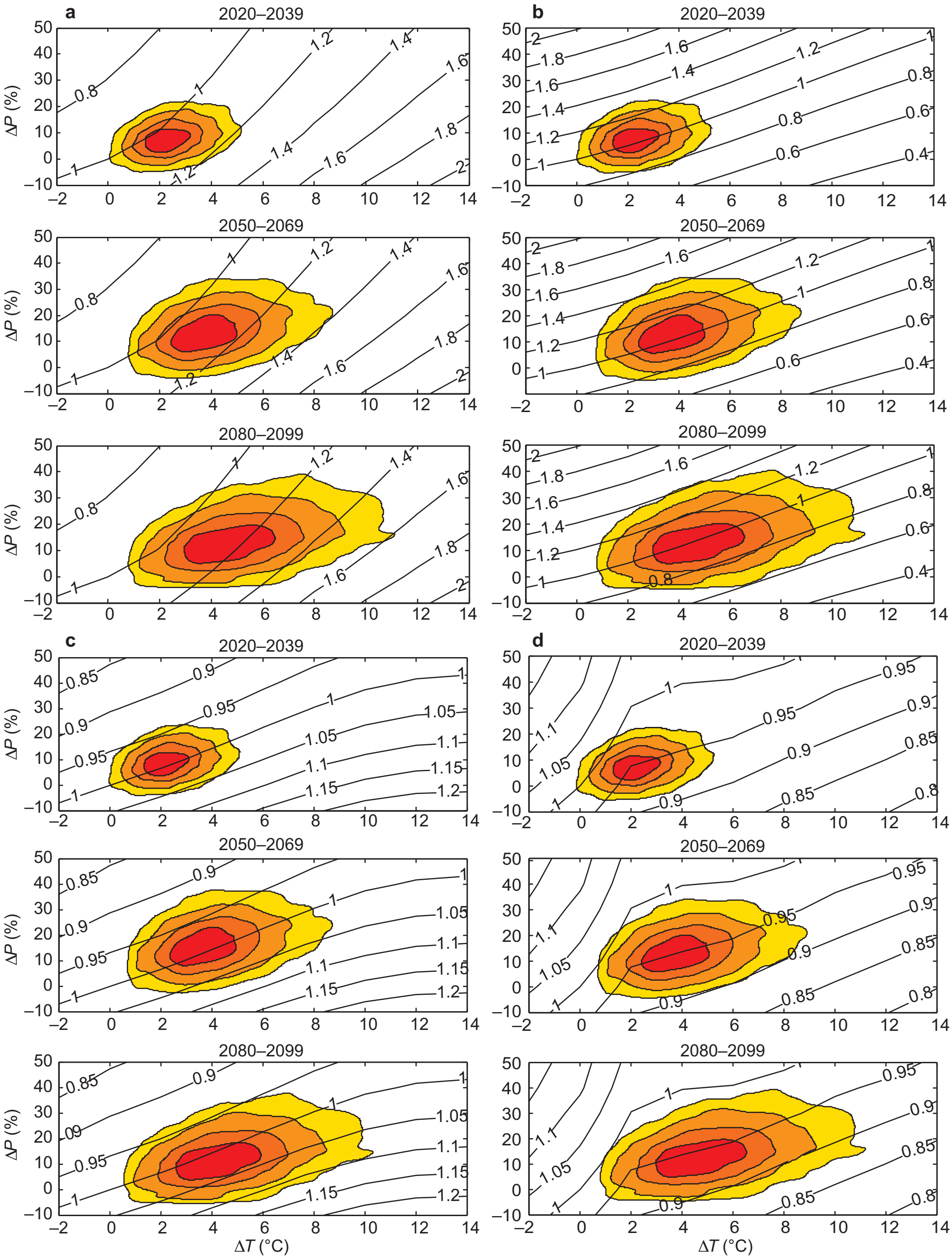

Fig. 6. Impact response surfaces (IRS) depicting the sensitivity to climate change of impact model results: (**a**) catchment soil moisture deficit; (**b**) annual runoff to lake; (**c**) DOC concentrations in influx to lake; (**d**) DOC concentrations in lake. The IRSs, drawn as labelled contour lines, represent the responses of the impact model chain to perturbations of *T* and *P*. The impact model results are expressed as standardized values where value 1 for no change in climate from 1961–1990 is the reference value for each variable given in Table 2; values larger than 1 mean an increase, and smaller than 1 a decrease in the model response; 90% of the climate change projections are within the outermost filled contour, 75% and 50% within the intermediate, and 25% within the innermost filled contour.

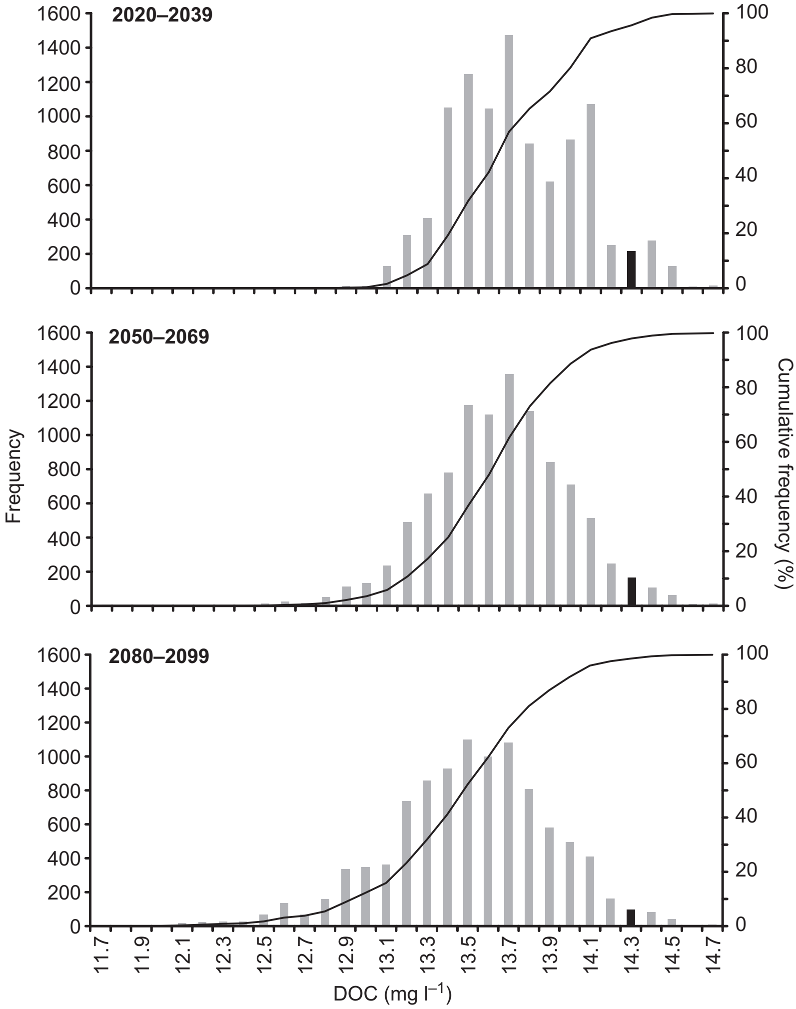

Fig. 7. Frequency and cumulative frequency distribution of 10 000 interpolations of modelled DOC in lake (mg l^{-1}) for three future time periods. The reference (14.3 mg l^{-1}) is set in black.

in the uppermost layer (0–1 m) of the lake will decrease towards the end of the century by at least 5%, and unlikely that they will decrease by more than 7%. The decrease in lake DOC is a consequence of drier soils, it is likely that the catchment soil will be at most 25% drier than at reference conditions (from 34 to 43 mm soil moisture deficit) and unlikely that it will be less than 12% drier (Table 2). In INCA-C, warmer soils produce more DOC. However, in the simulations presented here, warmer soils were also drier as a result of the greater HBV-estimated evapotranspiration. Thus, there was only a slight increase in the INCA-C estimated soil water DOC concentrations. The effect of this slight increase in concentrations combined with a slight decrease in runoff resulted in a very small estimated flux of DOC to the lake. Thus, the decrease in DOC in the lake can be linked to slightly less DOC export from the terrestrial environment as well as longer water residence times and greater opportunity for organic carbon processing in the lake.

Discussion

Frequency analysis of the probabilistic climate projections of *T* and *P* perturbations individually indicate a warming of 4 to 6 °C and an increase of 11% to 19% in precipitation for the likely-range (< 33% and > 66% probability). The joint frequency distribution, however, gives a broader range for the likely changes (Fig. 3). It is only in the light of the information inherent in the joint frequency distribution that quantitative estimates of the likelihood of impact responses can be made.

Laudon *et al*. (2012) suggested that climate change is likely to result in only modest increases in DOC concentrations in boreal streams. This is consistent with the soil-water DOC time series estimated from INCA-C, which showed only small increases. This small projected increase presented here is consistent with the simulations presented by Futter *et al*. (2009) in which the majority of the increase in modelled DOC concentrations between 1960 and 2100 was associated with the decline in atmospheric SO_4^{2-} deposition during the 1980s and 1990s. If this conclusion has general validity, it suggests that the large increases in surface water DOC concentrations observed near the end of the 20th century are unlikely to continue into the future. This should be contrasted with the large increases in lake DOC concentrations projected by Larsen *et al*. (2011) in a study of Norwegian lakes. Based on an empirical model using data from 1000 lakes, they suggested that a warmer climate would lead to increased terrestrial vegetation cover, which in turn will lead to increased DOC concentrations.

Climate change is expected to have profound effects on boreal forests including change in vegetation as well as increasing risk of fire and insect disturbance (Soja *et al*. 2007). Disturbance or forest management may lead to the replacement of the old growth spruce/pine forest at Valkea-Kotinen by other types of vegetation. A different vegetation community in the catchment might have very different patterns of evapotranspiration and litterfall than is observed at present, consequently different patterns of runoff and DOC export.

The projected changes in ice cover are consistent with the temporal pattern already observed in Valkea-Kotinen. Between 1990 and 2011, the duration of permanent ice cover has declined by approximately 15 days per decade (Jylhä *et al*. 2014). Model projections suggest that the decline of the duration of ice cover will continue at approximately the same rate throughout the 21st century.

The runoff and snow cover projections presented here are less extreme than those from earlier studies. Vehviläinen and Lohvansuu (1991) projected a 20%–50% increase in runoff and a complete loss of snow cover in southern Finland. Graham (2004) simulated a range of possible futures ranging from slight declines in runoff to a > 40% increase, depending on the GCM simulation used to drive the hydrological modelling. Veijalainen *et al*. (2010) projected decreases in the frequency of 100 year flood events across much of Finland by the end of the 21st century. This is consistent with the smaller spring runoff peaks simulated here as a result of less snow accumulation.

A number of simplifying assumptions were made in the model simulations presented here. The HBV model uses a degree-day index approach (Sælthun 1996) which may over-estimate evapotranspiration when the model is used for climate scenario projections (Lawrence and Haddeland 2011). However, there are significant uncertainties associated with the manner in which potential evapotranspiration is calculated in GCMs (Kingston *et al*. 2009). Teutschbein (2013) stated that potential evapotranspiration estimates from GCMs are strongly biased and not suitable for direct use in hydrological models. Furthermore, the current generation of RCM surface runoff estimates fail to satisfactorily reproduce present day runoff patterns (Teutschbein and Seibert 2010). In a study of 5 sites in Sweden, they showed that there are large discrepancies between monthly RCM estimated and observed runoff, suggesting there are problems with the manner in which surface runoff is represented in the current generation of climate models.

While the estimated runoff may be too low under some of the more extreme temperature scenarios, it has been suggested that this is most likely to be a problem during summer low flow periods (Lawrence and Haddeland 2011). Crossman *et al*. (2012) argued that this is not likely to be a problem in boreal and temperate systems where snowmelt or winter flows dominate the annual hydrography. As summer low flows typically accounted for 10% of annual simulated flows in the results presented here, discrepancies in the estimation of evapotranspiration is not likely to be a problem.

The MyLake simulations relied on time series of radiation, temperature, precipitation, relative humidity and wind speed. Only temperature and precipitation were perturbed when generating the response surfaces. This may have led to inconsistent estimates of relative humidity. While the HBV and INCA-C calibrations were quite successful, the MyLake calibration could well be improved and a new calibration might lead to somewhat different results for the impact sensitivities. We have only studied impact sensitivities directly related to projected changes in temperature and precipitation. A more comprehensive analysis would include sensitivities to model parameters and model structures. However, such an analysis would fail to represent the full uncertainty associated with effects of a changing climate on lake and catchment properties as it does not include the effects of fire, insect damage or changes in land management that can be expected in a warmer climate.

The IRS approach allows the impact modelling community to directly benefit from information inherent in the probabilistic climate projections. The probabilistic projections used here quantify uncertainties of the most important climate feedbacks and are constrained by different types of observations of the climate system (Harris *et al*. 2010). By the IRS approach we achieve a comprehensive quantification of the effect of climate uncertainty on impacts. Here, we presented the impact probabilities of nine catchment and lake variables, but the analysis can easily be extended to other variables simulated by the impact models, for example lake heat volume or chlorophyll concentrations. The approach is scenario-neutral (Prudhomme *et al*. 2010) and a major advantage of the IRS approach is that new climate scenarios can easily be incorporated by adding new points to the IRS.

A disadvantage of the IRS approach is that the change in variability is not treated. We used seasonal scalers to introduce seasonal variability into the daily time series of T and P to drive the impact models. The same seasonal scalers (calculated for 2000–2099) were used in all model simulations. By this simplification, we ignore the fact that seasonality varies over time in the climate projections (Harris *et al*. 2010).

MyLake (Saloranta *et al*. 2009, Forsius *et al*. 2010, Dibike *et al*. 2011), INCA-C (Aherne *et al*. 2008, Futter *et al*. 2009, Oni *et al*. 2012) and HBV (Lawrence and Haddeland 2011) were all used individually for simulating the effect of a changing climate on lake and catchment hydrological and biogeochemical processes. Using impact response surfaces obtained from a model chain of HBV, INCA-C and MyLake along with probabilistic climate projections can generate new insights that would not be possible with individual models or climate projections

Conclusions

We used the impact response surface approach to explore the impact of climate change on catchment and lake properties in Valkea-Kotinen. There is a clear seasonal pattern in the response of evapotranspiration and runoff to climate change. Probabilistic climate projections (Harris *et al*. 2010) were superimposed upon climate sensitivities of the model chain IRSs. Soil moisture deficit is sensitive to both temperature and precipitation change, and 12% to 25% drier soils are projected in response to the changing climate by the end of the century. The runoff from catchment to the lake is decreasing for some simulations and increasing in others. A larger decrease in runoff than –10% is unlikely, while it is likely that the increase is at most 5%. In our modelling analysis, drying soils were found to lead to slightly decreasing DOC concentrations in the uppermost layer (0–1 m) of the lake, at least by 5% from reference 14.3 mg C l^{-1} to 13.6 mg C l^{-1}, but unlikely to lower than 13.3 mg C l^{-1}. We found it likely that lake ice cover will continue to shorten, with a likely rate of about 1.5 days per year. The onset of lake stratification is likely to happen earlier, by 23 April, as compared with 12 May in reference conditions. The length of the stratified period is becoming longer, at a likely rate of about 0.5 days per year.

Despite the limitations in the structures, process descriptions and calibrations of our impact models and the uncertainties associated with the impacts of a changing climate on the boreal forest, we believe the impact response surface approach to be interesting, useful and informative. Using a model chain and impact response surfaces provides the means to visualize, quantify and communicate the likelihood of changes in the complex interactions of mass fluxes in boreal lakes and their catchments as a consequence of climate change.

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References

Aherne J., Futter M.N. & Dillon P.J. 2008. The impacts of future climate change and dissolved organic carbon on acidification recovery at Plastic Lake, Ontario. *Hydrol. Earth Syst. Sci.* 12: 383–392.

Arvola L., Rask M., Ruuhijärvi J., Tulonen T., Vuorenmaa J., Ruoho-Airola T. & Tulonen J. 2010. Long-term patterns in pH and colour in small acidic boreal lakes of varying hydrological and landscape settings. *Biogeochemistry* 101: 269–279.

*Arvola L., Salonen K., Keskitalo J., Tulonen T., Järvinen M. & Huotari J. 2014. Plankton metabolism and sedimentation in a small boreal lake — a long-term perspective. *Boreal Env. Res.* 19 (suppl. A): 83–96.

Bergström S. 1976. *Development and application of a conceptual runoff model for Scandinavian catchments*. SMHI Reports RHO no. 7, Norrköping, Sweden.

Bergström S. 1992. *The HBV model — its structure and applications*. SMHI Reports RH no. 4, Norrköping, Sweden.

Bergström I., Mäkelä K. & Starr M. (eds.) 1995. *Integrated Monitoring Programme in Finland. First National Report*. Ministry of the Environment, Environmental Policy Department, Helsinki, Finland.

Botev Z.I., Grotowski J.F. & Kroese D.P. 2010. Kernel density estimation via diffusion. *Annals of Statistics* 38: 2916–2957.

Curtis P.J. & Schindler D.S. 1997. Hydrologic control of dissolved organic matter in low-order Precambrian Shield lakes. *Biogeochemistry* 36: 125–138.

Crossman J., Futter M.N., Oni S.K., Whitehead P.G., Jin L. & Dillon P.J. 2012. Impacts of climate change on hydrology and water quality: future proofing management strategies in the Lake Simcoe watershed, Canada. *Journal of Great Lakes Research* 39: 19–32.

Dibike Y., Prowse T., Bonsal B., de Rham L. & Saloranta T. 2011. Simulation of North American lake-ice cover characteristics under contemporary and future climate conditions. *Int. J. Climatol.* 32: 695–709.

Einola E., Rantakari M., Kankaala P., Kortelainen P., Ojala A., Pajunen H., Mäkelä S. & Arvola L. 2011. Carbon pools and fluxes in a chain of five boreal lakes: a dry and wet year comparison. *J. Geophys. Res.* 116: 603009, doi:10.1029/2010JG001636.

Ford D.E. & Stefan H.G. 1980. Thermal predictions using integral energy model. *J. Hydraulic Div.* (*ASCE*) 106: 39–55.

Forsius M., Saloranta T., Arvola L., Salo S., Verta M., Ala-Opas P., Rask M. & Vuorenmaa J. 2010. Physical and chemical consequences of artificially deepened thermocline in a small humic lake — a paired whole-lake climate change experiment. *Hydrol. Earth Syst. Sci.* 14: 2629–2642.

Frieler K., Meinshausen M., Mengel M., Braun N. & Hare W. 2012. A scaling approach to probabilistic assessment of regional climate change. *J. Climate* 25: 3117–3144.

Fronzek S., Carter T.R., Räisänen J., Ruokolainen L. & Luoto M. 2010. Applying probabilistic projections of climate change with impact models: a case study for sub-arctic palsa mires in Fennoscandia. *Climatic Change* 99: 515–534.

Fronzek S., Carter T.R. & Luoto M. 2011. Evaluating sources of uncertainty in modelling the impact of probabilistic climate change on sub-arctic palsa mires. *Nat. Hazards Earth Syst. Sci.* 11: 2981–2995.

Futter M.N., Butterfield D., Cosby B.J., Dillon P.J., Wade A.J. & Whitehead P.G. 2007. Modelling the mechanisms that control in-stream dissolved organic carbon dynamics in upland and forested catchments. *Water Resour. Res.* 43, W02424, doi:10.1029/2006WR004960.

Futter M.N., Starr M., Forsius M. & Holmberg M. 2008. Modelling long-term patterns of dissolved organic carbon concentration in the surface waters of a boreal catchment. *Hydrol. Earth Syst. Sci.* 12: 437–447.

Futter M.N. & de Wit H.A. 2008. Testing seasonal and long-term controls of streamwater DOC using empirical and process-based models. *Sci. Total Environ.* 407: 698–707.

Futter M.N., Forsius M., Holmberg M. & Starr M. 2009. A long-term simulation of the effects of acidic deposition and climate change on surface water dissolved organic carbon concentrations in a boreal catchment. *Hydrology Research* 40: 291–305.

Futter M.N., Löfgren S., Köhler S.J., Lundin L., Moldan F. & Bringmark L. 2011. Simulating dissolved organic carbon dynamics at the Swedish Integrated Monitoring sites with the Integrated Catchments Model for Carbon, INCA-C. *Ambio* 40: 906–919.

Graham L.P. 2004. Climate change effects on river flow to the Baltic Sea. *Ambio* 33: 235–241.

Haaland S., Hongve D., Laudon H., Riise G. & Vogt R.D. 2010. Quantifying the drivers of the increasing colored organic matter in boreal surface waters. *Environ. Sci. Tech.* 44: 2975–2980.

Hammerschmidt C.R. & Fitzgerald W.F. 2006. Photodecomposition of methylmercury in an arctic Alaskan lake. *Environ. Sci. Tech.* 40: 1212–1216.

Harris G.R., Collins M., Sexton D.M.H., Murphy J.M. & Booth B.B.B. 2010. Probabilistic projections for 21st

* *Published in this issue*

century European climate. *Nat. Hazards Earth Syst. Sci.* 10: 2009–2020.

Hongve D., Riise G. & Kristiansen J.F. 2004. Increased colour and organic acid concentrations in Norwegian forest lakes and drinking water: a result of increased precipitation? *Aquat. Sci.* 66: 231–238.

Hondzo M. & Stefan H.G. 1933. Lake water temperature simulation model. *J. Hydraulic Eng.* 119: 1251–1273.

Hudson J.J., Dillon P.J. & Somers K.M. 2003. Long-term patterns in dissolved organic carbon in boreal lakes: The role of incident radiation, precipitation, air temperature, southern oscillation and acid deposition. *Hydrol. Earth Syst. Sci.* 7: 390–398.

*Jylhä K., Laapas M., Ruosteenoja K., Arvola L., Drebs A., Kersalo J., Saku S., Gregow H., Hannula H.-R. & Pirinen P. 2014. Climate variability and trends in the Valkea-Kotinen region, southern Finland: comparisons between the past, current and projected climates. *Boreal Env. Res.* 19 (suppl. A): 4–30.

Keller W. 2007. Implications of climate warming for Boreal Shield lakes: a review and synthesis. *Environ. Reviews* 15: 99–112.

Keskitalo J., Salonen K. & Holopainen A.-L. 1998. Long-term fluctuations in environmental conditions, plankton and macrophytes in a humic lake, Valkea-Kotinen. *Boreal Env. Res.* 3: 251–262.

Kingston D.G., Todd M.C., Taylor R.G., Thompson J.R. & Arnell N.W. 2009. Uncertainty in the estimation of potential evapotranspiration under climate change. *Geophys. Res. Lett.* 36(20), L20403, doi:10.1029/2009GL040267.

Korhonen J. 2002. Water temperature conditions of lakes and rivers in Finland in the 20th century. *The Finnish Environment* 566: 1–166. [In Finnish with English summary].

Larsen S., Andersen T. & Hessen D.O. 2011. Climate change predicted to cause severe increase of organic carbon in lakes. *Global Change Biology* 17: 1186–1192.

Laudon H., Buttle J., Carey S.K., McDonnell J., McGuire K., Seibert J., Shanley J., Soulsby C. & Tetzlaff D. 2012. Cross-regional prediction of long-term trajectory of stream-water DOC response to climate change. *Geophys. Res. Lett.* 39, L18404, doi:10.1029/2012GL053033.

Lawrence D. & Haddeland I. 2011. Uncertainty in hydrological modelling of climate change impacts in four Norwegian catchments. *Hydrol. Res.* 42: 457–471.

Lepistö A., Kortelainen P. & Mattsson T. 2008. Increased organic C and N leaching in a northern boreal river basin in Finland. *Global Biogeochem. Cycles* 22, GB3029, doi:10.1029/2007GB003175.

Leppäranta M. 1991. A review of analytical models of sea-ice growth. *Atmos. Ocean* 31: 123–138.

Mastrandrea M.D., Mach K.J., Plattner G.-K. Edenhofer, O., Stocker T.F., Field C.B., Ebi K.L. & Matschoss P.R. 2011. The IPCC AR5 guidance note on consistent treatment of uncertainties: a common approach across the working groups. *Climatic Change* 108: 675–691.

Monteith D.T., Stoddard J.L., Evans C.D., de Wit H.A., Forsius M., Høgåsen T., Wilander A., Skjelkvåle B.L., Jeffries D.S., Vuorenmaa J., Keller B., Kopácek J. & Vesely J. 2007. Dissolved organic carbon trends resulting from changes in atmospheric deposition chemistry. *Nature* 450: 537–541.

Nash J.E. & Sutcliffe J.V. 1970. River flow forecasting through conceptual models: Part I — A discussion of principles. *J. Hydrol.* 10: 282–290.

Oni S.K., Futter M.N., Molot L.A. & Dillon P.J. 2012. Modelling the long term impact of climate change on the carbon budget of Lake Simcoe, Ontario using INCA-C *Sci. Total Environ.* 414: 387–403.

Pirinen P., Simola H., Aalto J., Kaukoranta J.-P., Karlsson P. & Ruuhela R. 2012. *Climatological statistics of Finland 1981–2010*. Reports 2012:1, Finnish Meteorological Institute.

Posch M., Aherne J., Forsius M., Fronzek S. & Veijalainen N. 2008. Modelling the impacts of European emission and climate change scenarios on acid sensitive catchments in Finland. *Hydrol. Earth Syst. Sci.* 12: 449–463.

Prudhomme C., Wilby R.L., Crooks S., Kay A.L. & Reynard N.S. 2010. Scenario-neutral approach to climate change impact studies: application to flood risk. *J. Hydrol.* 390: 198–209.

Räisänen J. & Ruokolainen L. 2006. Probabilistic forecasts of near-term climate change based on a resampling ensemble technique. *Tellus* 58A: 461–472.

Rankinen K., Karvonen T. & Butterfield D. 2004. A simple model for predicting soil temperature in snow covered and seasonally frozen soil: model description and testing. *Hydrol. Earth Syst. Sci.* 8: 706–716.

Rask M., Verta M., Korhonen M., Salo S., Forsius M., Arvola L., Jones R.I. & Kiljunen M. 2010. Does lake thermocline affect methyl mercury concentrations in fish? *Biogeochemistry* 101: 311–322.

Riley M.J. & Stefan H.G. 1988. MINLAKE: a dynamic lake water quality simulation model. *Ecol. Model.* 43: 155–182.

*Ruoho-Airola T., Hatakka T., Kyllönen K., Makkonen U. & Porvari P. 2014. Temporal trends in the bulk deposition and atmospheric concentration of acidifying compounds and trace elements in the Finnish Integrated Monitoring catchment Valkea-Kotinen during 1988–2011. *Boreal Env. Res.* 19 (suppl. A): 31–46.

Saarsalmi A., Starr M., Hokkanen T., Ukonmaanaho L., Kukkola M., Nöjd P. & Sievänen R. 2007. Predicting annual canopy litterfall production for Norway spruce (*Picea abies* (L.) Karst.) stands. *Forest Ecol. Manage.* 242: 578–586.

Sælthun N.R. 1996. The "Nordic" HBV Model. *Description and documentation of the model version developed for the project Climate Change and Energy Production*. NVE Publication 7, Norwegian Water Resources and Energy Administration, Oslo.

Saloranta T. 2000. Modeling the evolution of snow, snow ice and ice in the Baltic Sea. *Tellus* 52A: 93–108.

Saloranta T.M. & Andersen T. 2007. MyLake — a multi-year lake simulation model suitable for uncertainty and sensitivity analysis simulations. *Ecol. Model.* 207: 45–60.

Saloranta T., Forsius M., Järvinen M. & Arvola L. 2009. Impacts of projected climate change on the thermodynamics of a shallow and deep lake in Finland: model simulations and Bayesian uncertainty analysis. *Hydrol.*

* *Published in this issue*

Res. 40: 234–247.

Schindler D.W., Bayley S.E., Parker B.R., Beaty K.G., Cruickshank D.R., Fee E.J., Schindler E.U. & Stainton M.P. 1996. The effects of climatic warming on the properties of boreal lakes and streams at the experimental lakes area, Northwestern Ontario. *Limnol. Oceanogr*. 41: 1004–1017.

Schindler D.W., Curtis P.J., Bayley S.E., Parker B.R., Beatty K.G. & Stainton M.P. 1997. Climate-induced changes in the dissolved organic carbon budget of boreal lakes. *Biogeochemistry* 36: 9–28.

Snucins E. & Gunn J. 2000. Interannual variation in the thermal structure of clear and colored lakes. *Limnol. Oceanogr.* 45: 1639–1648.

Soja, A.J., Tchebakova N.M., French N.H.F., Flannigan M.D., Shugart H.H., Stocks B.J., Sukhinin A.I., Parfenova E.I., Chapin F.S.III & Stackhouse P.W.Jr. 2007. Climate-induced boreal forest change: predictions versus current observations. *Global and Planetary Change* 56: 274–296.

Starr M. & Ukonmaanaho L. 2001. Results from the first round of the integrated monitoring soil chemistry subprogramme. In: Ukonmaanaho L. & Raitio H. (eds.), *Forest condition in Finland*, National Report 2000, Finnish Forest Research Institute, Helsinki, Research Papers 824, pp. 140–157.

Starr M. & Ukonmaanaho L. 2004. Levels and characteristics of TOC in throughfall, forest floor leachate and soil solution in undisturbed boreal forest ecosystems. *Water Air Soil Pollut. Focus* 4: 715–729.

Starr M., Saarsalmi A., Hokkanen T., Merilä P. & Helmisaari H.-S. 2005. Models of litterfall production for Scots pine (*Pinus sylvestris* L.) in Finland using stand, site and climate factors. *Forest Ecol. Manage*. 205: 215–225.

Teutschbein C. 2013. *Hydrological modeling for climate change impact assessment*. Ph.D. thesis in Physical Geography at Stockholm University.

Teutschbein C. & Seibert J. 2010. Regional climate models for hydrological impact studies at the catchment scale: a review of recent modeling strategies. *Geography Compass* 4: 834–860.

Ukonmaanaho L., Starr M. & Ruoho-Airola T. 1998. Trends in sulphate, base cations and H^+ concentrations in bulk precipitation and throughfall at integrated monitoring sites in Finland 1989–1995. *Water Air Soil Pollut*. 105: 353–363.

Vähätalo A.V., Salonen K., Münster U., Järvinen M. & Wetzel R.G. 2003. Photochemical transformation of allochthonous organic matter provides bioavailable nutrients in a humic lake. *Archiv Hydrobiol*. 156: 287–314.

Vähätalo A.V., Aarnos H. & Mäntyniemi S. 2010. Biodegradability continuum and biodegradation kinetics of natural organic matter described by the beta distribution. *Biogeochemistry* 100: 227–240.

Vehviläinen B. & Lohvansuu J. 1991. The effects of climate change on discharges and snow cover in Finland. *Hydrol. Sci. J.* 36: 109–121.

Veijalainen N., Lotsari E., Alho P., Vehviläinen B. & Käyhkö J. 2010. National scale assessment of climate change impacts on flooding in Finland. *J. Hydrol*. 391: 333–350.

Verta M., Salo S., Korhonen M., Porvari P., Paloheimo A. & Munthe J. 2010. Climate induced thermocline change has an effect on the methyl mercury cycle in small boreal lakes. *Sci. Total Environ*. 408: 3639–3647.

Vuorenmaa J., Forsius M. & Mannio J. 2006. Increasing trends of total organic carbon concentrations in small forest lakes in Finland from 1987 to 2003. *Sci. Total Environ*. 365: 47–65.

*Vuorenmaa J., Salonen K., Arvola L., Mannio J., Rask M. & Horppila P. 2014. Water quality of a small headwater lake reflects long-term variations in deposition, climate and in-lake processes. *Boreal Env. Res*. 19 (suppl. A): 47–65.

Weiss M. 2011. Future water availability in selected European catchments: a probabilistic assessment of seasonal flows under the IPCC A1B emission scenario using response surfaces. *Nat. Hazards Earth Syst. Sci*. 11: 2163–2171.

Wetterhall F., Graham L.P., Andréasson J., Rosberg J. & Yang W. 2011. Using ensemble climate projections to assess probabilistic hydrological change in the Nordic region. *Nat. Hazards Earth Syst. Sci*. 11: 2295–2306.

Ylhäisi J.S., Tietäväinen H., Peltonen-Sainio P., Venäläinen A., Eklund J., Räisänen J. & Jylhä K. 2010. Growing season precipitation in Finland under recent and projected climate. *Nat. Hazards Earth Syst. Sci*. 10: 1563–1574.

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