Interaction of impacts of doubling CO$_2$ and changing regional land-cover on evaporation, precipitation, and runoff at global and regional scales

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ABSTRACT: The Community Climate System Model version 2.0.1 is running for 40 years under 355 ppm CO$_2$ conditions, without and with natural and anthropogenic land-cover changes that are assumed in the inner core of four hydrothermally different, but similar-sized ($=3.27 \times 10^6$ km$^2$) regions (Yukon, Ob, St. Lawrence, Colorado, and lands adjacent to them). A further set of simulations assumes 710 ppm CO$_2$ conditions without and with these land-cover changes. Impacts of (1) doubled CO$_2$, (2) changed land-cover, and (3) the interaction between doubled CO$_2$ and changed land-cover on the four regional water cycles are elucidated using analysis of variance plus multiple testing. For the Yukon, Ob, and St. Lawrence regions, doubling CO$_2$ significantly increases precipitation, evapotranspiration, and residence time nearly year-round; the opposite is true for precipitation and evapotranspiration in Colorado. In general, doubling CO$_2$ slows down water cycles regardless of land-cover changes. Since land-cover changes occur locally, they more strongly affect regional than global water cycling. Sometimes land-cover changes alone reduce regional-scale precipitation and evapotranspiration. Water-cycle changes of comparable absolute magnitude can occur in response to either changed land-cover or doubled CO$_2$. Significant interactions between the two treatments indicate that local land-cover changes, even if they have little impact under reference climate conditions, may have substantial regional impact in a warmer climate. Increased residence time after doubling CO$_2$ indicates a generally increased influence of upwind regions on downwind regions. If land-cover changes occur concurrently with CO$_2$ changes, they will have farther-reaching impact than under reference CO$_2$ conditions. Thus, due to atmospheric transport the interaction between impact of land-cover changes and CO$_2$ doubling on water-cycle-relevant quantities may occur even in regions with unchanged land-cover. A sensitivity study for tripled CO$_2$ showed similar results, but more pronounced slowed-down regional water cycles and interaction of the two treatments. Copyright © 2008 Royal Meteorological Society

KEY WORDS: ANOVA; CCSM; CO$_2$; interaction of anthropogenic impacts; land-cover changes; multiple testing

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1. Introduction

Increasing greenhouse-gas concentrations and land-cover changes may each affect weather and climate, and hence the atmospheric water cycling at various scales (e.g. Cotton and Pielke, 1995). Evapotranspiration (sum of sublimation, evaporation, and transpiration) and cloud and precipitation formation are essential water-cycle processes. Obviously, increasing CO$_2$ and changing land-cover affect water cycles through different mechanisms. Increased CO$_2$ affects these processes indirectly, as absorbed radiation warms the atmosphere. Any phase-transition processes, however, depend non-linearly on temperature. Land-cover changes affect evapotranspiration directly via altered plant physiological parameters (e.g. stomatal resistance, root length, leaf area index (LAI), shielding of the ground, interception-storage capacity). Land-cover changes also influence evapotranspiration indirectly: (1) modified aerodynamic roughness alters wind speed, and (2) modified radiative properties (e.g. albedo, emissivity) change energy partitioning at the surface–atmosphere interface. Thus, the altered exchanges of matter, heat, and momentum cause changes in cloud and precipitation formation.

Recently, many studies have examined the impact of changed greenhouse-gas concentrations (e.g. Mitchell, 1989; Houghton et al., 1996, 2001; Yang et al., 2003) or changed land-cover (e.g. Sud et al., 1996; Chase et al., 1999; Bounoua et al., 2002; Molders and Kramm, 2007) on water-cycle-relevant quantities (e.g. evapotranspiration, precipitation, runoff). A general conclusion from studies of water-cycle responses to increased CO$_2$ is that globally averaged precipitation, evapotranspiration, and precipitable water are increasing (e.g. Roads et al., 1996; Douville et al., 2002; Weatherald and Manabe, 2002). Several studies (e.g. Douville et al., 2002; Bosilovich et al., 2005) indicate a ‘slowed-down global water cycle’ due to increased residence time; (defined as the ratio of
precipitable water to precipitation, Trenberth, 1998). Residence time is a measure of atmospheric water storage. However, changes in globally averaged evapotranspiration, precipitation, and residence time are not equally distributed, but rather vary on regional scales (e.g. Douville et al., 2002; Bosilovich et al., 2005). Therefore, there is an urgent need to explore regional water-cycle complexity and response to global changes, and also to compare regional responses to each other.

Recent investigations using regional weather forecast or climate models showed that mid-latitude and high-latitude land-cover changes may appreciably alter near-surface energy budgets and cloud and precipitation distribution in the region of, and adjacent to land-cover changes (e.g. Chase et al., 1999; Bounoua et al., 2000; Mölders and Olson, 2004). General Circulation Model (GCM) studies show that regional land-cover changes, such as the Amazon deforestation, can alter global and/or regional water cycles at mid-/high latitudes by modifying large-scale atmospheric circulation (e.g. Sud et al., 1996; Zhang et al., 1996; Chen et al., 2001; Avisar and Werth, 2005).

In nature, land-cover and CO2 change concurrently; however, few studies have examined the combined effect of these changes on specific regions or globally. In mid-latitudes, for instance, conversion of natural vegetation to agricultural land may compensate for warming caused by rising greenhouse gases and increasing regional precipitation (e.g. Stohlgren et al., 1998; Bounoua et al., 2002). Owing to complex non-linear relationships among water-cycle-relevant processes, it is difficult to assess the concurrent impact of two changes based on knowledge of the impacts of individual changes (e.g. Mölders, 2000).

Therefore, we examine how the combined effects of doubling CO2 and changing land-cover affect water-cycle-relevant quantities globally in four selected similarly-sized regions. The Yukon, Ob, St. Lawrence, Colorado, and adjacent lands (Figure 1) are selected as study regions because of their different thermal and hydrological regimes, and different position with respect to moisture sources (oceans) and the large-scale circulation pattern. The main aim of our study is to examine whether there are regional differences in the responses to land-cover changes under different CO2 conditions. Special foci are on whether assumed land-cover changes impact water-cycle-relevant quantities in a similar manner: (1) under different conditions of CO2, (2) in different regions, (3) in different seasons, and (4) whether the combined impacts of doubling CO2 and changing land-cover are diminished or enhanced compared to their individual effects. Answering these questions requires separating and quantifying the variance induced by the different treatments (changed land-cover, doubled CO2 concentrations) and by the interaction between these treatments so that the treatments can be compared objectively. A method that permits such separation and quantification is ANalysis Of VAriance (ANOVA) (Montgomery, 1976; von Storch and Zwiers, 1999).

The ANOVA-technique has been successfully used to answer a variety of atmospheric science questions. Zwiers (1996), for instance, applied one- and two-way ANOVA to simulate and observe geopotential height, temperature, and stream-function fields. He found that boundary conditions add to the potential predictability of variance for 500-hPa geopotential, 850-hPa temperature, and 300-hPa stream-function at the inter-annual scale. Rovell (1998) applied ANOVA to a six-member ensemble of 45-year climate simulations performed with identical sea-surface temperature, but different initial atmospheric conditions to assess seasonal predictability of precipitation based on oceanic forcing. By applying ANOVA to stream-functions and temperature fields of simulations with different CO2 levels and land-cover, Zhao et al. (2001) found that the interaction of responses to CO2 and land-cover changes affects the large-scale circulation pattern and the transports of temperature perturbations related to altered land-cover. By using an unreplicated two-way ANOVA design to partition the variance of simulated precipitation for Fairbanks, Alaska into four

Figure 1. Location of the study regions: Yukon (Y), Ob (O), St. Lawrence (S), and Colorado (C). The area of each region is about 3.27 × 10^6 km^2. Note that due to the map projection, high-latitude study regions appear larger than mid-latitude study regions.
treatments (urbanization, release of aerosols, heat, and moisture) plus any combination of the four, Mölders and Olson (2004) found statistically significant impacts of aerosol release, moisture, and combined aerosol release and urbanization on downwind precipitation.

In our study, we apply ANOVA to data obtained from four simulations with the Community Climate System Model (CCSM) version 2.0.1 (Kiehl and Gent, 2004). These simulations alternately assume 355 ppm and 710 ppm CO2 concentration, without and with land-cover changes in the centre of the four study regions. The ANOVA serves to objectively analyse individual and combined impacts of both changed land-cover and increased CO2 concentrations on the global water cycle and on regional water-cycle-relevant quantities of the four similar-sized study areas. The ANOVA design uses an F-test at the 95% confidence level to identify significant changes in, and interaction between the impacts of the treatments on water-cycle-relevant quantities. To avoid 'significance by chance' (type I errors), we apply the Tukey HSD (honest for significant differences) multi-testing (e.g. Pearson and Hartley, 1972). To further test the robustness of changes, ANOVA plus multiple testing is also applied to results from simulations with tripled CO2 without and with land-cover changes in the centre of the four study regions.

2. Experimental design

2.1. Model description and initialization

CCSM consists of four model components: The Climate Atmosphere Model (CAM) version 2, the Parallel Ocean Program (POP), the Common Land Model (CLM) version 2, and the Community Sea Ice Model (CSIM) version 4.0.1. A flux coupler exchanges data between individual model components without flux correction (Kiehl and Gent, 2004).

The CAM (Kiehl and Gent, 2004) is the successor to the Atmosphere General Circulation Model: Community Climate Model version 3 (AGCM: CCM3). It applies hybrid vertical coordinates with the terrain-following \( \sigma \) – coordinate merged into pressure coordinates around 100 hPa. Major changes from CCM3 to CAM include formulations for cloud condensed water, cloud-fraction and overlap, and long-wave water vapour absorptivity and emissivity (Kiehl and Gent, 2004). CAM simulates deep convection by a plume-ensemble approach, in which convective available potential energy is removed from a grid-column at an exponential rate in accord with Zhang and McFarlane (1995). Local convection and large-scale dynamics may interact through pressure perturbations caused by cloud-momentum transport (Zhang et al., 1998). Shallow convection is treated based on Hack (1994) and Zhang et al. (1998). At the resolvable scale, cloud and precipitation formation processes are parameterized by a bulk-microphysical scheme in accord with Rasch and Kristjánsson (1998), which takes into account water-vapour condensation and associated temperature changes, condensate evaporation, and condensate-to-precipitation conversion.

CLM (Dai et al., 2003) considers ten layers each for soil temperature, soil water and ice content, one vegetation layer, and (depending on snow depth) up to five snow layers. Every grid-cell is divided into four land-cover types (glacier, lake, wetland, vegetation). Vegetation is further divided into as many as four plant functional types (PFT). Precipitation can be intercepted by the canopy and evaporate from interception storage. CLM considers the topographic control of runoff by a TOPographically based rainfall-runoff MODEL (TOPMODEL)-like approach (e.g. Dai et al., 2003; Niu et al., 2005). Runoff parameterization considers surface and subsurface flow outside wetlands, and base flow in wetlands (Dai et al., 2003).

The Los Alamos National Laboratory’s POP (Smith et al., 1992) is used to simulate ocean processes. This ocean model comprises a North Pole displaced from the Arctic Ocean up to Greenland; thus, it requires no filtering of the ocean solution for the Arctic. This feature is of great advantage in our study: The resulting fine resolution (\(< 1^\circ\)) guarantees an open Bering Strait and Northwest Passage and therefore permits more realistic representation of the Arctic halocline (Holland, 2003).

The CSIM (Briegleb et al., 2004; Holland et al., 2006) considers subgrid-scale ice-thickness distribution with five ice categories. Each category can cover a fractional area within a grid-cell. Consequently, CSIM can more realistically simulate ice conditions and thermodynamic processes over the Arctic Ocean than sea-ice models without these features. Including subgrid-scale sea-ice-thickness heterogeneity permits us to capture sea-ice retreat in response to a warmer climate.

CCSM is integrated into a fully coupled mode with 26 vertical layers and a spectral truncation of 42, corresponding to a horizontal grid increment of about 2.8°×2.8°. CCSM is initialized with the astrophysical conditions for 1 January, 1990. Each model component is spun-up in its off-line mode and all simulations begin from the same spun-up condition.

2.2. Simulations and study regions

Many climate change studies use transient CO2 scenarios, ensembles thereof or ensembles of a given transient CO2 scenario from different GCMs (e.g. Kiehl et al., 2006; Meehl et al., 2006; Tebaldi et al., 2006). To better understand processes and interactions between land-cover and CO2 impacts and to avoid misinterpreting signals caused by model differences or signals resulting from constantly increasing CO2 concentration, we perform the following four 40-year simulations: 355 ppm CO2 concentration, current land-cover (CTR); 710 ppm CO2 concentration, current land-cover (DBL); 355 ppm CO2 concentration, altered land-cover (LUC); and 710 ppm CO2 concentration, altered land-cover (LUCDBL). Furthermore, for a sensitivity study, simulations with tripled CO2 without and with land-cover changes in the centre of the
four study regions are performed to examine the robustness of changes. Since inserting land-cover changes and increasing CO₂, as well as altering both, slightly disturbs equilibrium conditions in the various compartments, the first 10 simulation years are discarded as spin-up time to achieve a new equilibrium (e.g. Figure 2).

Four similar-sized (≈3.27 × 10⁶ km²) study regions are chosen so that we can easily compare the magnitude of regional changes. Global climate models have, almost without exception, indicated an amplification of global warming in the polar regions in greenhouse experiments (e.g. Houghton et al., 1990; Teng et al., 2006); due to various feedback mechanisms, changes in the water cycle due to altered radiative forcing should be visible first in the polar regions and then propagate to other regions of the world. Therefore, two polar study regions, namely the Yukon and Ob basins and areas adjacent to them, are selected. These regions differ with respect to elevation, wetness, and the ocean into which they discharge. The Yukon River flows to the Bering Sea through complex terrain strongly underlain by cold and discontinuous permafrost under conditions of the present relatively dry climate, while the Ob discharges into the Arctic Ocean after flowing through relatively flat, wet terrain underlain by warm and discontinuous permafrost that is less extensive than the permafrost of the Yukon basin. The Ob basin is chosen to represent the conditions of reduced permafrost that are expected to occur in response to global warming. In mid-latitudes, water vapour and precipitation are mainly transported into a region by frontal systems that may be modified by local surface characteristics (e.g. Loose and Bornstein, 1977). Most climate models predict an increase in winter precipitation for mid-latitude regions in response to doubling CO₂ (e.g. Cubasch et al., 1995). For continental-size regions located far away from the oceans or in (semi)arid regions the amount of precipitation resulting from local recycling of water gains importance (e.g. Eltahir and Bras, 1996). To investigate the altered behaviour of these different types of dominance in the water-cycle pathways,

Figure 2. Differences in near-surface air temperature between conditions of doubled CO₂ (DBL) and control (CTR), changed land-cover (LUC) and CTR, and concurrently changed land-cover and doubled CO₂ (LUCDBL) and CTR used to determine spin-up time. The zero line is bold.

The dotted areas indicate positive differences greater than 2 K, the shaded areas indicate negative differences exceeding 1 K.
the mid-latitude St. Lawrence and Colorado regions are selected. Their most obvious differences are terrain (moderate vs. mountainous) and climate (humid vs. semi-arid). The latter represents an area of limited water availability under current climate conditions where the human presence is still expanding (see US Census, 2006). The former is an example of an economically developed area with, on average, sufficient available water.

Modelled land-cover changes are realized by altering the PFT in each region’s inner core. Note that if within a grid-cell, a percentage of PFT A changes to B in one place and an area of the same size covered by B changes to A, the grid-cell will not experience a net change in PFT according to the mosaic approach used in CLM. Such kinds of changes alter landscapes and their heterogeneity (cf. Mölders, 2000), but not their fractional PFT cover (cf. Dai et al., 2003). We orient the assumed land-cover changes at a rate typically found in these regions of the world on time scales of 30 years or so (Bork, pers. comm. 2000). Wildfires have disturbed high-latitude boreal forests for millennia (e.g. Stocks et al., 1994). We orient the assumed land-cover changes at a rate typically found in these regions of the world on time scales of 30 years or so (Bork, pers. comm. 2000). Wildfires have disturbed high-latitude boreal forests for millennia (e.g. Stocks et al., 1994) with increasing area in the last 30 years (Podur et al., 2002) and even a doubling in North America in the last 20 years (Stocks et al., 2000). Therefore, we reduced the percentage fraction of coniferous forest by 7.6 and 7.0% in the inner core of the Yukon and Ob regions and correspondingly increased the percentage coverage of grassland, the typical succession landscape (DNR, 2000). Subsidization policies aim at increasing agricultural use in Alaska and Siberia. Thus, we increased the percentage of crops in the inner core of the Yukon and Ob regions by 2.1 and 5.5% at the cost of grassland. These conversions reduce roughness length, interception-storage capacity, and LAI, and modify albedo and emissivity. Additionally, in summer, plant-available water is taken from levels closer to the surface as grass/crops have shorter roots than trees. Under typical high-latitude winter conditions, the assumed land-cover changes yield increased albedo, because unlike tall vegetation, short vegetation is totally snow-covered.

In the inner core of the St. Lawrence region, some deciduous forest is being replaced by grass (15.5%) or crops (6.8%). To represent the land abandonment occurring in these areas (e.g. Bonan, 2001), we also convert 16.1% of cropland to either grassland (6.8%) or forest (9.3%).

In a warmer climate, irrigation needs will increase. However, one can assume that with decreased water resources, but increased demands for food production lower yields than today may become profitable. Under these circumstances, to maintain at least the same average harvest, the acreage of agriculturally used land may increase (Bork, pers. comm. 2000). Thus, conversion of grassland to cropland is the main land-cover change (4.5%) assumed for the inner Colorado region. Since wildfires frequently occur in this region (e.g. Westerling et al., 2003) 1.8% of forest is changed to grassland. The persistent droughts around the millennium reduced the cattle-herd sizes. Since shrubs establish in non-grazed areas, 14.5% of the grassland is converted to shrubs. Conversion from grass to crops/shrubs and vice versa mainly affects albedo.

2.3. Analysis

We determine climatologies from the reference simulation (CTR) for evaluating and investigating the impacts of the various treatments (Figure 3). Data used in the evaluation include monthly precipitation from the Global Precipitation Climatology Centre (GPCC) for 1971 to 2000; near-surface air temperature (SAT) from the European Re-Analysis 40 (ERA40) (Uppala et al., 2005) for 1971 to 2000; and cloud-fraction data based on the International Satellite Cloud Climatology Project (ISCCP; Rosow and Schiffer, 1999) for 1984 to 2004. These data sets have a resolution of 2.5° × 2.5° and are interpolated to the 2.8° × 2.8° resolution of CTR. Skill scores (root mean square error (RMSE), bias, standard deviation of error (SDE), accuracy, Heidke skill-score (HS) are calculated in accord with Anthes (1983), Anthes et al. (1989), and Wilks (1995). In addition, we summarize results from existing CCSM evaluations and inter-comparison studies.

We perform a 2 × 2-factorial design, single-block, fixed-effects ANOVA to analyse impacts of changing land-cover, doubling CO₂, and the interaction between the two treatments. In the term 2², the exponent represents two factors (land-cover changes and doubled CO₂), and the base represents two levels (treatment switched on or off). To examine the robustness of changes, ANOVA is applied using the data from the simulations with tripled CO₂ conditions in a sensitivity study.

In general, any simulated variable (e.g. precipitation) can be decomposed into an overall mean, individual effects, interactions between effects, and error terms. In terms of the sum of squares (SS), a variable’s variation can be written as (e.g. Montgomery, 1976)

$$SS_T = SS_A + SS_B + SS_{AB} + SSE$$

where $SS_T$, $SS_A$, $SS_B$, $SS_{AB}$, and $SSE$ are the sum of squares of the total T, factor A, factor B, the second-order interaction AB, and error E, respectively. A factor’s sum of squares divided by its degree of freedom provides that factor’s mean square which can be used to evaluate the forcing’s impact (Montgomery, 1976). For a 2²-factorial design, multiple replicates are required to determine the degree of freedom and the mean square of the error.

In this study, each simulation represents one treatment, i.e. in Equation (1) A, B, and AB represent the impact of doubling CO₂, changing land-cover, and the interaction between the impacts of the two, respectively. Each year’s data for a given month at either a single grid-point or averaged over a region are taken as a replicate for that month. For example, January’s precipitation $P$ for each year at a single grid-point for each simulation is taken as a replicate for January. Thirty replicates of January precipitation from each simulation are summed as $P_{CTR}$.
Figure 3. Annual cycles of monthly averaged: (a) evapotranspiration (ET), (b) precipitation (P), (c) runoff (RO), and (d) residence time (RT) for the globe (grey), the Yukon (solid), Ob (dashed), St. Lawrence (dotted), and Colorado (dash-dotted) regions.

The level of the two factors, changed land-cover and doubled CO$_2$, considered here are denoted as $I$ and $J$, respectively. Since there are two possibilities, treatment ‘switched on’ or ‘switched off’, both $I$ and $J$ take the value of 2 in our design. The degree of freedom of the factors $A$, $B$, and their second-order interaction $AB$ are $I - 1$, $J - 1$ and $(I - 1) \cdot (J - 1)$, respectively.

With independent replicates, the factor A mean square divided by the error mean square will follow an F-distribution; degrees of freedom are given by $I - 1$ (for the mean square of factor A) and $I \cdot J \cdot N - 1$ (for the error). The lag-one auto-correlations $\rho$ of these replicates (e.g. precipitation) are $\approx 0.15$, indicating that replicates are not completely independent of each other. Therefore, the error degree of freedom is calculated using an equivalent sample size in accord with von Storch and Zwiers (1999):

$$N_{eq} = \frac{N}{1 + \sum_{n=1}^{N} \left(1 - \frac{1}{N}\right) \rho}$$

Thus, with the equivalent sample size $N_{eq} = 25$ the error degree of freedom is 96.
An F-test is applied to measure a factor’s quantitative importance, and to detect significance. The F-test’s null hypothesis assumes no significant impact of changed land-cover, doubled CO\textsubscript{2}, or treatment interactions on water-cycle-relevant quantities. To reject the null hypothesis at the 95% confidence level, the F-value must exceed 3.94. We perform a post-ANOVA Tukey HSD test (hereafter denoted Tukey test) to exclude significance by chance. This test determines a new critical value that serves to evaluate whether changes between any two pairs of means reach significance. The Tukey test is applied to the annual cycle time series as well as to the grid-cell by grid-cell significance testing. Worldwide this post-ANOVA multiple testing rejects variances at the 95% confidence level in 36% (74%) or fewer locations for doubling CO\textsubscript{2} in December-January-February (DJF) [June-July-August (JJA)], and in about 70% of the locations for land-cover changes. Most of the rejections are in the Southern Hemisphere and in the Arctic Ocean. In our study regions, the Tukey test rejects significance in less than 10 and 75% of the cases for doubling CO\textsubscript{2} and altering land-cover, respectively. The high percentage of rejection within the study regions may be explained by the low number of points for which a significant impact of land-cover change is detected by ANOVA. Note that for the American south-west region, precipitation in Alaskan precipitation and PNA index only marginally correlate (e.g. Leathers et al., 1991).

3. Model performance

3.1. Global evaluation

Near-SAT plays an essential role in determining evapotranspiration. CCSM slightly overestimates the global SAT (Table I); the underestimated cloud-fraction and lower terrain height in the model than in nature contribute to this positive bias. Bias is highest in March (2 K) and lowest in June (1 K). The overall model performance for SAT is better in DJF, JJA, and September-October-November (SON) than in March-April-May (MAM). The warm bias can be explained by simulated sensible (latent) heat flux that is too high (low) during snowmelt, warming the lower atmosphere at northern high latitudes (cf Bonan et al., 2002; Kiehl and Gent, 2004).

Compared to various published evapotranspiration climatologies (e.g. Croley et al., 1998; Su et al., 2006) and ERA40 data (Uppala et al., 2005) CCSM accurately captures the spatial pattern and annual cycle of evapotranspiration for the globe and the four study regions (Li et al., 2007).

Compared to Rossov et al. (1996) 1983–2001, 3-h cloud observations, CCSM overestimates cloud-fraction by up to 20% in the Tropics, but underestimates it by up to 30% in the extra-tropics (Dai and Trenberth, 2004). On the basis of Rossov and Schiffer’s (1999) more recent data, we find that CCSM underestimates global cloud-fraction in all months and on annual average (Table I). The negative bias is greatest (up to 5% absolute) during boreal summer. The premature onset of deep convection (Dai and Trenberth, 2004) leads to systematic errors in cloud-fraction and precipitation in all study regions (Table I; Figure 4).

CCSM was evaluated within the Program for Climate Model Diagnosis and Intercomparison (PCMDI) reports (Meehl et al., 2000; Covey et al., 2003) that compared precipitation simulated by 18-coupled GCMs with observations. Differences between simulated and observed global annual precipitation range between −0.1 and +0.4 mm/day, while correlations between simulated and observed precipitation patterns range between 0.7 and 0.9 mm/day. CCSM’s spatial correlation coefficient and bias equals 0.79 and 0.08 mm/day (= 2.4 mm/mon; cf Table I), respectively. According
Table I. Simulated and observed annual means, root means square errors (RMSE), bias, and standard deviation of errors (SDE) of near-surface air temperature (SAT), cloud-fraction (CF), and precipitation (P) for the globe and the study regions.

<table>
<thead>
<tr>
<th>Simulated mean</th>
<th>Observed mean</th>
<th>RMSE</th>
<th>Bias</th>
<th>SDE</th>
</tr>
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<tr>
<td>SAT (°C)</td>
<td>Global</td>
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<td>5.5</td>
<td>3.3</td>
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<tr>
<td></td>
<td>Yukon</td>
<td>−6.3</td>
<td>−6.9</td>
<td>3.2</td>
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<td></td>
<td>Ob</td>
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<td>−0.8</td>
<td>3.1</td>
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<tr>
<td></td>
<td>St. Lawrence</td>
<td>8.2</td>
<td>6.5</td>
<td>2.8</td>
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<tr>
<td></td>
<td>Colorado</td>
<td>16.9</td>
<td>13.4</td>
<td>4.5</td>
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<tr>
<td>CF (%)</td>
<td>Global</td>
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<td>67</td>
<td>22</td>
</tr>
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<td></td>
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<td></td>
<td>Ob</td>
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<td>21</td>
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<tr>
<td></td>
<td>Colorado</td>
<td>42</td>
<td>54</td>
<td>18</td>
</tr>
<tr>
<td>P (mm/month)</td>
<td>Global</td>
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<td>61.7</td>
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<td>41.9</td>
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<td>26.3</td>
</tr>
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</table>

to the PCMDI, CCSM’s precipitation biases (precipitation overestimation, southward-shifted South Pacific convergence zone, excessive northern mid-latitude precipitation) are common in coupled GCMs (e.g. Johns et al., 1997; Meehl et al., 2000; Covey et al., 2003; Furevik et al., 2003).

For all study regions, precipitation SDE and RMSE are highest during summer (Figure 4). CCSM acceptably captures regional-scale annual averages and spatial precipitation patterns in the Yukon, Ob, and St. Lawrence regions (Li et al., 2007). However, a large discrepancy exists between Colorado regional simulated and observed annual precipitation cycles (see Section 3.5 for discussion).

Accuracy and HS are determined for global and regional precipitation thresholds of 5, 10, 15, 30, 45, 75, 100, 150, and 300 mm/month and 20, 30, 40, 50, 60, 70, 80, 90, and 100 mm/month, respectively. CCSM simulates global precipitation with an accuracy >78% for all thresholds in all seasons (e.g. Figure 4). CCSM precipitation simulations exceed 60% accuracy in all the study regions and show the highest accuracy in winter.

The HS evaluates precipitation against random values (cf. Wilks, 1995). Global HS-values are around 0.6 for precipitation thresholds <150 mm/month. HS decreases substantially for thresholds >150 mm/month, indicating that CCSM does not accurately capture heavy tropical precipitation. HS-values range from 0.3 to 0.6 for the Yukon region, while they are less than 0.4 for all thresholds for the Ob, St. Lawrence, and Colorado regions (Figure 4). Note that HS between 0.2 and 0.3 are typical for good performance on the regional scale (e.g. Zhong et al., 2005).

CCSM-projected runoff agrees acceptably with observed runoff data from the University of New Hampshire Global Runoff Data Center (Niu et al., 2005). Overall, CCSM produces global mean runoff within observed ranges (e.g. Dai et al., 2001; Bonan et al., 2002).

3.2. Yukon

CCSM shows warm bias during winter and on annual average, and cold bias during summer. SDE and RMSE are higher in the cold than in the warm season (Table I). The errors may be related to the parameterizations of surface upward long-wave radiation, snow-age, and albedo and to biome classification (see also Bonan et al., 2002).

CCSM overestimates cloud-fraction (3–11% absolute) for most months and on annual average (Table I). On the basis of a relatively low RMSE, bias, and SDE, one can conclude that CCSM accurately captures the Yukon region cloudiness.

CCSM systematically overestimates precipitation (Table I) because the elevation of the Alaska Range and Brooks Range is much lower in the model than in nature, allowing cyclones to move farther north in the model than they do in nature. Since 45–50% of the precipitation days are trace events in the northern part of the Yukon region (e.g. Yang et al., 1998) reported observations may be lower than the real values (e.g. Li et al., 2007). Thus, trace precipitation also causes systematic errors. Higher errors are found in summer and early fall when convective precipitation plays a stronger role than in other months (Figure 4).

3.3. Ob

For most months, CCSM overestimates SAT (Table I). Overall, CCSM represents SAT better in the cold season than in the warm season. Errors in simulated cloud-fraction are the main cause of incorrect SAT. CCSM slightly overestimates (underestimates) cloud-fraction during the cold (warm) season, and underestimates cloud-fraction on annual average (Table I).

CCSM captures the north–south gradient of annual precipitation in the Ob region, but slightly overestimates precipitation on annual average (Table I). The positive bias is highest in October (12 mm/month), and SDE and
RMSE are highest in July at 15.5 and 17.5 mm/month, respectively (Figure 4).

3.4. St. Lawrence

Warm bias is found for St. Lawrence year-round (Table I). The highest bias occurs in March (2.6 K) and the lowest in May and November (<1 K). In contrast to the Ob region, in the St. Lawrence region CCSM performs better in the warm (especially in early summer and in fall) than in the cold season; SDE and RMSE are about 1 K higher in winter than in summer. The reason is that average temperature difference between the ocean and the continent is generally greater in winter (7.2 K) than in summer (−0.6 K). Therefore, differences between modelled and natural land–sea distribution cause greater errors in winter than in summer.

CCSM underestimates cloud-fraction on annual average (Table I). During the warm season, cloud formation by local convection (e.g. over the Great Lakes) is underestimated because of the premature onset of convection.
Of all four study regions, CCSM has the poorest performance for cloud-fraction in the St. Lawrence region. CCSM underestimates precipitation year-round (cf. Figure 4; Table I). As discussed by Li et al. (2007), errors in predicted precipitation can be related to the representation of the Great Lakes, discrepancies in land–sea distribution between the model and nature, and shortcomings of the convection scheme.

3.5. Colorado

Simulated SAT show a warm bias of 3.5 K on annual average (Table I). RMSE and bias have similar annual cycles with a July maximum of 8 and 9 K, respectively. SDE is also highest (4 K) during summer. In semi-arid regions, large systematic errors in summer temperature can be related to the excessive emission of long-wave radiation from the surface simulated by CCSM (cf. Bonan et al., 2002). Furthermore, errors also occur because of the differences between modelled and natural terrain heights.

CCSM has appreciable difficulties in simulating the cloud and precipitation formation in the mountainous terrain of the Colorado region. CCSM underestimates the Colorado region cloud-fraction substantially during summer. In summer, bias, RMSE, and SDE can be as much as 24, 21, and 32% (absolute cloud-fraction), respectively. These underestimations in summer cloud-fraction contribute to the underestimation of precipitation and result in a simulated water cycle that is drier than that found in nature. Note that CCSM overestimates precipitation in most months of the other seasons. Patterns of increasing annual precipitation from the West to the East of the region exist both in the GPCC and CTR-data-derived climatology. However, the low elevation of the Rocky Mountains in the model leads to misrepresentation of the annual cycle of observed precipitation. The discrepancy in precipitation partly results from the too early onset of convection simulated by CCSM (Dai and Trenberth, 2004) and the fact that measured mountain precipitation may not accurately represent total precipitation (e.g. Dingman, 1994). However, precipitation simulation over elevated terrain is a common problem for GCMs (Johns et al., 1997; Flato et al., 2000; Coquard et al., 2004) and mesoscale models (Colle et al., 2000; Nara-pusetty and Mölders, 2005; Zhong et al., 2005) if placed in the context of other models’ performances. CCSM’s precipitation simulations represent current state-of-the-art. Thus, any interpretation of results carries with it a higher uncertainty for the Colorado than for the other study regions.

3.6. General remarks

Modern GCMs show 1–3 K climate sensitivity in response to doubled CO₂ (e.g. Cubasch et al., 2001; AchutaRao et al., 2004; Kiehl et al., 2006). CCSM’s climate sensitivity of 2.3 K is within this range (Kiehl and Gent, 2004). In CCSM, globally averaged near-SATs increase 1.2 and 1.8 K in boreal summer and winter, respectively, in response to doubling CO₂. Generally, projected warming is greater at high latitudes than at mid-latitudes because of the snow-albedo feedback. CCSM’s air-temperature increase projections agree with results from other GCMs (e.g. Mahfouf et al., 1994; Cubasch et al., 2001; Achuta Rao et al., 2004; Kiehl et al., 2006).

According to existing evaluations of CCSM presented by us and others (e.g. Meehl et al., 2000; Dai et al., 2001; Covey et al., 2003; Dai and Trenberth, 2004; Kiehl and Gent, 2004; Niu et al., 2005; Li et al., 2007), this model is well suited for investigating water-cycle-relevant questions. Our study focuses on differences and interactions between impacts, rather than absolute values; we can assume that the CCSM shortcomings discussed above will affect these processes similarly in the various simulations.

4. Results and discussion

4.1. Global water cycle

The ANOVA plus multiple testing shows that for more than 34% of the earth’s surface, doubling CO₂ significantly impacts evapotranspiration during DJF (Figure 5(a) and (c)). Many of these areas are located at high latitudes. During JJA, the total area with significant evapotranspiration changes in response to doubled CO₂ is smaller than in DJF (Figure 5(b) and (e)). In JJA, doubling CO₂ hardly affects mid-latitude evapotranspiration, but significantly increases evapotranspiration north of 60°N and in the Tropics. Increased high-latitude evapotranspiration results from the appreciable increase of near-SAT (≤4.3 K in DBL, and ≤3.7 K in LUCDBL) in response to doubling CO₂. Less sea-ice forms in these relatively warmer climate regimes; in consequence, the Arctic Ocean can supply more water vapour to the atmosphere in DBL and LUCDBL than in CTR. This sea-ice retreat yields increased high-latitude late fall/early winter evapotranspiration in the pan-Arctic.

According to the ANOVA plus multiple testing, land-cover changes and the interaction between the two treatments significantly affect evapotranspiration in a smaller area during boreal winter than during other seasons (Figure 5(d) and (f)) for the following reasons. During spring (MAM), summer (JJA), and fall (SON), vegetation is active in most regions of the Northern Hemisphere. Thus, differences between original and new vegetation in plant physiological, radiative, and aerodynamic properties (e.g. LAI, interception-storage capacity, soil-water uptake by roots, roughness length, albedo emissivity) yield different near-surface fluxes of momentum, latent andensible heat, and net radiation, with impacts on water cycle and state variables. In boreal winter, vegetation is inactive. In snow-covered areas, taller vegetation (coniferous forest, deciduous forest) rises above the snow when shorter vegetation (grassland, cropland) is already snow-covered. Under these conditions, the assumed land-cover changes affect evapotranspiration indirectly by differences in

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Figure 5. Seasonally averaged evapotranspiration under control (CTR) conditions for (a) December-January-February (DJF) and (b) June-July -August (JJA). Differences in seasonally averaged evapotranspiration between conditions of doubled CO₂ (DBL)-CTR during (c) DJF and (e) JJA, and concurrently changed land-cover and doubled CO₂ (LUCDBL)-CTR during (d) DJF and (f) JJA. Stippled areas indicate significant changes (ANOVA plus multiple testing; ≥95% confidence level) due to doubling CO₂ and the interaction of changed land-cover and doubled CO₂. There are a few significant changes for LUC (therefore not shown). Note that the ANOVA plus multiple testing indicates significant interaction between the effects of changing land-cover and doubling CO₂ on evapotranspiration, not significant changes in evapotranspiration.

roughness length and surface-radiative parameters. The change from taller to shorter vegetation increases near-surface wind speed, and therefore potential evapotranspiration (note that this impact strongly depends on wind speed). During a snow event trees intercept snow, thereby increasing albedo. As the interception storage empties, albedo decreases again. Consequently, conversion from taller to shorter vegetation means that once a closed snow-cover is established, albedo remains relatively high; more incoming radiation is reflected back from the modified than the original landscape. Therefore, if high-latitude land-cover changes as assumed here, near-SAT will decrease once snow-cover forms (snow-albedo feedback). Lower temperatures reduce potential evapotranspiration and (over-)compensate for the effect of reduced roughness length on potential evapotranspiration for most near-surface wind speeds. Altered albedo and emissivity affect net radiation, with consequences for partitioning between sensible and latent heat flux (evapotranspiration times latent heat consumption). In summary, land-cover changes indirectly contribute to lower winter near-SATs via snow-albedo feedback.

Since land-cover changes only occur in four locally limited areas, land-cover changes alone only marginally affect the global annual evapotranspiration cycle (Figure 6), but this cycle increases significantly in response to doubled CO₂ or to concurrent doubled CO₂ and changed land-cover. In the latter case, significant global-scale change occurs in response to increased CO₂ rather than to changed land-cover; the ANOVA reveals no significant interaction between the two treatments on that scale. However, doubling CO₂, changing land-cover, or combining the two may either enhance or reduce precipitation via cloud-evaporation feedback. More evapotranspiration promotes cloud and precipitation formation. Increased cloudiness decreases evapotranspiration by reducing short-wave radiation reaching the surface; in response, cloud formation is reduced,
increasing evapotranspiration again. Previous precipitation may trigger this effect. Obviously, cloud-evaporation feedback is highly non-linear because of various processes involved at different scales.

Changes in globally averaged annual precipitation and evapotranspiration (Figure 6) exhibits correlations >0.863, 0.897, and 0.935 for DBL-CTR, LUC-CTR, and LUCDBL-CTR, respectively. According to the ANOVA plus multiple testing, doubling CO\textsubscript{2} significantly increases DJF precipitation in northwestern China, Mongolia, western Europe, along the equator, and north of 60° N (Figure 7(a) and (c)). In the latter region, increased atmospheric moisture supply resulting from sea-ice retreat is a main reason. In JJA, the area experiencing significant precipitation changes in response to doubling CO\textsubscript{2} is smaller than in winter, especially at mid-latitudes and high latitudes (Figure 7(b) and (e)).

The ANOVA plus multiple testing shows that the impact of land-cover changes alone and the impact of the interaction between the two treatments on precipitation are much weaker than the impact of doubling CO\textsubscript{2} alone (Figure 7(d) and (f)). Nevertheless, and although land-cover changes occur at local (not continental) scales, the interaction between increased CO\textsubscript{2} and changed land-cover significantly affects precipitation in regions far from the areas of altered land-cover (Figure 7(d) and (f)).

Globally averaged precipitation decreases in LUC boreal summer. However, from May to October, LUCDBL precipitation increases exceed DBL increases, indicating that land-cover changes affect global water cycling differently under reference and enhanced CO\textsubscript{2} conditions (Figure 6(b)). In LUC, reduced LAI via deforestation causes decreased transpiration leading to less summer precipitation. In LUCDBL, ≈1.3 K greenhouse effect warming and increased surface wind due to reduced roughness length (a deforestation consequence) yield enhanced evapotranspiration (Figure 6(a)) compared to CTR, DBL, and LUC. Note that as compared to CTR, insignificant changes in winter-snow amount (<4.5 mm decrease in the Yukon, <4.5 mm increase in Ob, St. Lawrence and Colorado regions for DJF) in DBL, LUC, and LUCDBL are not strong enough to show any effect on the peaks of snowmelt and runoff in spring for the study regions.

In response to doubling CO\textsubscript{2}, runoff increases significantly in Alaska, Canada, the northern continental USA,
most of Europe, and central and south Africa during DJF (Figure 8(a) and (c)). In JJA, significant impact of doubling CO$_2$ on runoff occurs in some river basins close to the coasts (Figure 8(b) and (e)). As expected, local land-cover changes hardly affect globally averaged runoff (Figure 6(c)). Land-cover changes and the interaction of changed land-cover and doubled CO$_2$ have little effect on winter or summer runoff over most of the globe (Figure 8(d) and (f)).

If both treatments are applied concurrently globally averaged runoff changes significantly according to the $t$-test, except in February, March, June, and July. The ANOVA plus multiple testing reveals significant interaction between the two treatments in October, when runoff increases more in LUCDBL than after doubling CO$_2$ alone. Interestingly, under reference CO$_2$ conditions land-cover changes reduce October runoff (Figure 6(c)) due to the snow-albedo effect. In LUC, decreased near-surface temperature reduces evapotranspiration and precipitation, creating a positive feedback that finally reduces runoff. However, in LUCDBL, under doubled CO$_2$ temperature increases substantially (1.8 K) causing enhanced evapotranspiration and precipitation, while land-cover changes decrease LAI, reducing the amount of water intercepted by and evaporated from the canopy, and enhancing LUCDBL runoff compared to DBL; consequently, the treatment interaction enhances LUCDBL runoff more than the effect of either treatment alone.

A region’s residence time is defined as the ratio of regionally averaged monthly precipitable water to precipitation. Increased residence time means that all phases of water remain longer in the atmosphere before reaching the ground. Thus, water is more likely to be transported farther from its source than in the reference condition. The opposite is true for decreased residence time. On average, doubling CO$_2$ significantly increases globally averaged residence time, but land-cover changes have marginal impact (Figure 6(d)). No significant interaction exists when the two treatments are applied concurrently. Consequently, the significant (by $t$-test) LUCDBL residence-time increase results from doubling CO$_2$, and not from changing land-cover or from non-linear interactions between the treatments.

Besides sea-ice retreat, snow-albedo feedback, and cloud-evaporation feedback, other mechanisms modify regional water cycles. Land-cover changes assumed in...
Figure 6. Differences in global average monthly (a) evapotranspiration (ΔET), (b) precipitation (ΔP), (c) runoff (ΔRO), and (d) residence time (ΔRT). The thin solid, thin dashed, and thick solid lines represent conditions of doubled CO₂ (DBL) - control (CTR), changed land-cover (LUC)-CTR, and concurrently changed land-cover and doubled CO₂ (LUCDBL)-CTR, respectively. Circles and triangles indicate significant impact of doubling CO₂ and changing land-cover, respectively. Squares identify significant interaction between the impacts of changing land-cover and doubling CO₂. Significance refers to ANOVA plus multiple testing. Stars indicate significant differences between LUCDBL and CTR according to a Student’s t-test. Note that no significant changes occur for LUC.

this study, for instance, decrease interception storage. Since summer near-SATs differ only marginally between LUC and CTR, additional water contributes to infiltration and runoff rather than to surface evaporation. In the following discussion, we will call this feedback loop ‘consequences of reduced interception’ for simplicity. Furthermore, as discussed below, slight large-scale circulation shifts as documented by altered PNA and NAO indices may alter moisture advection and hence affect regional water cycles tremendously.

The ANOVA plus multiple testing including tripled CO₂ provides similar results as for doubled CO₂. However, the interaction between tripled CO₂ and changed land-cover is more pronounced and the regional water cycles slow down more than for doubled CO₂. On the basis of the sensitivity study, we may conclude that these changes are the most robust.

4.2. Yukon

The ANOVA plus multiple testing reveals a significant interaction between doubled CO₂ and altered land-cover for November (Figure 9(a)) evapotranspiration. November evapotranspiration increases in both DBL and LUCDBL because of the sea-ice retreat. However, DBL evapotranspiration increases ≈3 mm/month more than LUCDBL evapotranspiration because of the decreased LUCDBL temperature resulting from the snow-albedo feedback over altered land-cover.

In most months, precipitation is affected more by doubling CO₂ than by changing land-cover (Figure 9(b)). According to the ANOVA plus multiple testing, DBL precipitation significantly increases in most months except for July, September, and October, primarily due to increased evapotranspiration and moisture advection in response to higher SAT. This increase of DBL winter moisture advection mainly results from the increased near-SAT and evaporation rather than from the stronger pressure gradient (Li et al., 2007). The winter PNA indices in DBL and LUCDBL marginally differ from the PNA index in CTR (Figure 10(a)); marginal changes of the Aleutian Low occur in response to doubled CO₂. Thus, changes in winter precipitation closely relate to changes in the PNA index. This result agrees with Leathers et al. (1991) findings on the correlation between winter precipitation and PNA index in the American Northwest. The impact of land-cover changes on Yukon...
winter precipitation is small. Their impact on winter PNA index is mixed with slightly more years, but with less magnitude of positive PNA anomalies. According to the ANOVA plus multiple testing, Yukon region land-cover changes affect the precipitation similarly under different CO2 conditions.

The impact of changed land-cover on runoff is marginal because most runoff occurs during snowmelt when the Yukon vegetation is only marginally active (Figure 9(c)). Note that conifers still actively photosynthesize at temperatures around the freezing point. In DBL and LUCDBL, relatively higher temperatures under doubled CO2 retard fall snow-cover onset, and hasten spring snowmelt. The slight shift in favour of more liquid precipitation also contributes to higher winter runoff and lower runoff during snowmelt than in CTR (Figure 9(c)). In the northern Yukon, snowmelt occurs around late May/early June. Consequently, runoff increases significantly due to increased CO2 only, indicating that under either reference or doubled CO2 conditions, land-cover changes have a marginal impact on the runoff.

According to the ANOVA plus multiple testing, land-cover changes significantly increase the November residence time due to a significantly decreased November precipitation (Figure 9(d)). September residence time increases less in LUCBL than in DBL, and the treatment interaction is significant. This diminished increase results from reduced evapotranspiration due to the consequences of reduced interception in LUCDBL as explained earlier, and counteracts increased potential evaporation caused by higher temperatures under increasing CO2. Consequently, the September atmosphere contains less precipitable water in LUCDBL than in DBL.

4.3. Ob

In summer, the Ob region’s upper soil is wet; soil moisture fractions (ratio of actual volumetric water content to porosity) exceed 77% for all projections. Land-cover changes yield decreased evapotranspiration from
late spring to late fall (Figure 11(a)). The original forest takes up water for transpiration from deep in the active layer; the modified landscape’s grass and crops draw water from shallower levels. Evapotranspiration significantly increases year-round in DBL due to generally warmer conditions. In LUCDBL, evapotranspiration also increases, but less than in DBL except in August. The smaller evapotranspiration increase in LUCDBL rather than in DBL suggests that the impact of land-cover changes on evapotranspiration mitigates the effect of doubling CO2. Nevertheless, LUCDBL evapotranspiration increases significantly according to a t-test except for July, September, and October. Despite opposing impacts of changed land-cover and doubled CO2 on summer evapotranspiration (decrease vs. increase), the interaction between changed land-cover and increased CO2 will remain insignificant if these changes occur simultaneously.

The ANOVA plus multiple testing identifies strongly decreased LUC September precipitation, exceeding the decrease after doubling CO2 or after combined treatments (Figure 11(b)). As pointed out earlier, the assumed land-cover changes decrease the evapotranspiration. In September, winds from the south-west prevail leading to advection of air masses from the North Atlantic and Mediterranean Sea. Since the land-cover changes in the St. Lawrence region are in the upwind of the Atlantic, the NAO index is analysed. No obvious change in LUC of the NAO index is found in September. This means that the strong decreases in LUC September precipitation cannot be explained by a modified NAO index and is related to the reduced local recycling of precipitation.

In September, warmer air advected across the western and southern Ob regional boundaries in LUCDBL combined with marginally increased sensible heat fluxes yield a 0.8 K temperature increase as compared to CTR. Increased temperature and decreased moisture supply to the atmosphere result in reduced relative humidity (Figure 12), increased residence time, and less precipitation than in CTR (Figure 11(b) and (d)).

Concurrently doubling CO2 and changing land-cover increase precipitation significantly (t-test) in February, May, June, November, and December, but according
to the ANOVA, no significant treatment interaction exists. Since this increased precipitation is mainly due to increased evapotranspiration in a warmer climate and since evapotranspiration decreases in response to land-cover changes, we attribute increased LUCDBL evapotranspiration and the resulting increased precipitation to increased CO2 (Figure 11(b)). As discussed above, in September, however, land-cover changes alone strongly reduce precipitation because they increase sensible heat fluxes at the cost of evapotranspiration, thereby significantly increasing residence time (Figure 11(a) and (d)).

During April and May, runoff decreases by ≤7.5 mm/month in DBL, but increases by 3 mm/month in LUC despite decreased LUC precipitation in May (Figure 11(b) and (c)). If both doubled CO2 and changed land-cover occur, the effect of reduced interception loss on runoff will partly compensate for decreased snowmelt resulting from reduced winter-snow accumulation under increased CO2. Thus, in the Ob region, land-cover changes alter precipitation partitioning between various regional water-cycle pathways during snowmelt for both CO2 concentration levels.

In LUC, the snow-albedo feedback yields a cooler atmosphere than in CTR once the snow-cover is established. Since saturated water-vapour pressure exponentially decreases with decreasing temperature, LUC residence time decreases slightly in winter, and significantly in September and October (Figure 11(d)). In contrast, warmer DBL temperatures increase the residence time. For concurrent changed land-cover and doubled CO2, counteracting treatment effects reduce the winter increase of residence time, but during summer and fall, the combined impact enhances the magnitude of residence-time increase compared to DBL. Although the ANOVA detects no significant treatment interactions for these months, the t-test shows that increased LUCBDL residence time is significant in July, September, and October. The increased summer and fall residence time in DBL, LUC, and LUCBDL can be explained by changes in temperature and precipitation. In September, for instance, the residence time increase (≤2.2 days) peaks in all three simulations due to (up to 2 K) increased air temperatures. This substantial temperature increase accompanies a relatively small decrease in evapotranspiration (2 mm/month).
Figure 8. As in Figure 5, but for the runoff. Note that contour spacing for (a) and (b) differs from that in Figures 5(a) and (b) and 7(a) and (b), and that contour spacing for (c) to (f) differs from Figures 5(c)–(f) and 7(c)–(f) for better illustration.

4.4. St. Lawrence

Since the St. Lawrence region is relatively humid with considerable available water, atmospheric demand mainly governs evapotranspiration. Therefore, temperature increases associated with doubling CO$_2$ (1.2–2.9 K compared to CTR) dominate changes in regional evapotranspiration. DBL evapotranspiration significantly increases in all months with a maximum of 4.5 mm/month in April (Figure 13(a)). Land-cover changes affect evapotranspiration insignificantly under reference CO$_2$. In contrast, under doubled CO$_2$, especially in March and October, land-cover changes further enhance evapotranspiration increase caused by doubling CO$_2$, while LUC evapotranspiration is reduced (Figure 13(a)). The ANOVA plus multiple testing indicates significant (90% confidence level) interactions between changed land-cover and doubled CO$_2$. In March, cloud-evaporation feedback explains the different impact of land-cover changes found for LUC and LUCDBL. In October, the decreased evapotranspiration under LUC as compared to CTR results from the snow-albedo feedback. However, despite the fact that LUCDBL albedo is slightly higher than DBL albedo, increased cloudiness and reduced long-wave radiation loss in LUCDBL (Figure 14) considerably enhance the temperature increases in DBL.

Impacts of land-cover changes on precipitation are small and insignificant year-round in the St. Lawrence region under reference CO$_2$ conditions (Figure 13(b)). However, in response to the land-cover changes, the winter NAO index decreases resulting in more years with more pronounced negative phases of NAO (as seen in Figure 10(b)). During winter, the St. Lawrence region is governed by prevailing western wind directions, and the major moisture source is advection of moist air from the Great Lakes area. Thus, the shift towards negative NOA index means a weakening of the Westerly
prevailing winds, colder conditions, decreased storminess and reduced precipitation in the St. Lawrence region. If land-cover changes and CO2 doubles at the same time, the land-cover change impact will partially offset the significant precipitation increases occurring in response to doubled CO2 (Figure 13(b)). Runoff changes marginally in response to land-cover changes or to the interaction between treatments (Figure 13(c)). Similar to the Ob region, changed land-cover alone or doubled CO2 alone have opposite impacts on the runoff during snowmelt in the St. Lawrence region. However, concurrent treatments decrease spring runoff more in LUCDBL than in DBL, mainly due to enhanced evapotranspiration and often lower precipitation in LUCDBL than DBL (Figure 13(a) and (b)).

Doubling CO2 increases residence time in most months; land-cover changes do the opposite under reference CO2 conditions (Figure 13(d)). The two treatments significantly interact in June; land-cover changes enhance the response to doubled CO2, but decrease residence time under reference CO2. In June, evapotranspiration increases more in LUCDBL than in DBL due to cloud-evaporation feedback. The exponential relationship between temperature and saturation water-vapour pressure increases the amount of water vapour that can be taken up before saturation is reached and precipitation formation starts. The greater near-surface wind speed in LUCDBL than in DBL helps to export moisture from the region before precipitation reaches the ground (Figure 13(b)). Therefore, residence time increases in LUCDBL compared to DBL (Figure 13(d)).

4.5. Colorado

In the Colorado region, evapotranspiration is commonly limited by soil moisture under reference CO2 and land-cover change conditions. Thus, even if warmer conditions after doubling CO2 permit higher potential evapotranspiration than under reference CO2, both DBL and LUCDBL evapotranspiration will decrease from summer to fall compared to CTR (Figure 15). Nevertheless, changing land-cover decreases June evapotranspiration by 4 mm/month, four times the decrease after doubling CO2. Reduced canopy storage, and new vegetation that

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Figure 8. (Continued).

takes water from levels closer to the surface than the original vegetation explain the appreciably decreased LUC June evapotranspiration (Figure 15(a)).

In accord with Leathers et al. (1991) PNA and precipitation only marginally correlate for the Colorado region. Thus, as expected, any changes in PNA (Figure 10) cannot explain changes in Colorado precipitation (therefore not further discussed), i.e. the changes must have more local causes. The annual cycle of the impact of land-cover changes on Colorado precipitation closely follows the annual cycle of impact on evapotranspiration with a correlation coefficient of 0.875 (Figure 15(a) and (b)). The same is true for land-cover change effects under doubled CO$_2$ ($r = 0.837$). The significant September DBL precipitation decrease ($-6$ mm/month) results from increased surface sensible heat fluxes ($\leq 1.6$ W/m$^2$), but decreased evapotranspiration ($-3$ mm/month). Thus, energy is partitioned differently in DBL than in CTR. Owing to enhanced sensible heat fluxes, near-SAT is 2.9 K higher in DBL than in CTR. Atmospheric moisture demand increases, but is not satisfied by evapotranspiration.

September evapotranspiration is hardly affected by land-cover changes (Figure 15(a)), but precipitation increases due to increased moisture advection. Doubling CO$_2$ further increases moisture advection; the increase is greater in the combined treatment than in LUC alone, but less than in DBL. Precipitation decreases similarly in LUCDBL and DBL; land-cover changes affect precipitation differently under reference and doubled CO$_2$ conditions. Obviously, doubling CO$_2$ decreases evaporation, i.e. the fraction of precipitation originating from within the region is enhanced by land-cover changes under reference, but not under doubled CO$_2$ conditions.

The ANOVA indicates that doubling CO$_2$ significantly decreases runoff in February, March, August, and September, but neither the impact of changing land-cover alone or the effect of treatment interaction is significant (Figure 15(c)). DBL runoff decreases because more precipitation falls as rain in winter, winter runoff increases, less snow accumulates, and the snowmelt runoff is reduced. Decreased precipitation and enhanced residence time in summer and early fall significantly decrease DBL August and September runoff (Figure 15(b)–(d)).

From July to October changing land-cover reduces residence time (by up to 1.8 days) (Figure 15(d)), but doubling CO$_2$ significantly increases residence time (by

Figure 9. Differences in the annual cycles of domain-averaged monthly (a) evapotranspiration (ΔET), (b) precipitation (ΔP), (c) runoff (ΔRO), and (d) residence time (ΔRT) for Yukon. The thin solid, thin dashed, and thick solid lines represent conditions of doubled CO₂ (DBL) – control (CTR), changed land-cover (LUC) – CTR, and concurrently changed land-cover and doubled CO₂ (LUCDBL) – CTR, respectively. Circles and triangles indicate significant impact of doubling CO₂ and changing land-cover, respectively. Squares identify significant interaction between the impact of changing land-cover and doubling CO₂. Significance refers to ANOVA plus multiple testing. Stars indicate significant differences between LUCDBL and CTR according to a Student’s t-test.

Figure 10. Time series of winter (a) Pacific North America (PNA) and (b) North Atlantic Oscillation (NAO) indices. The gray lines, thin solid black lines, thin dashed black lines and thick solid black lines stand for reference (CTR) conditions, doubled CO₂ (DBL), changed land-cover (LUC), and concurrently changed land-cover and doubled CO₂ (LUCDBL), respectively.

5. Conclusions
The fully coupled CCSM version 2.0.1 is used to examine the response of the global water cycle and the water cycles of four regions, the Yukon, Ob, St. Lawrence, and Colorado and lands adjacent to them, to doubling CO₂ (DBL), land-cover changes that only occur in the centre of these regions (LUC), and the same land-cover changes under doubled CO₂ conditions (LUCDBL). An ANalysis Of VAriance with F-test and Tukey’s test at the 95% confidence level serves to evaluate treatment influences on
Regional water cycles slow down under doubled CO\(_2\) conditions both without and with land-cover changes, i.e. residence time increases. Increased residence time suggests that upwind conditions can influence regional water cycles; a region’s effect extends further downwind than under reference CO\(_2\) conditions, or to relate this in a different way, a region is affected by more distant upwind regions.

Independent of CO\(_2\) conditions, land-cover changes affect water-cycle-relevant quantities only slightly on the global scale, because they are assumed to occur only in the core of the four study areas. Nevertheless, notable impacts of land-cover change alone and significant impacts of treatment interactions are not restricted to areas of changed land-cover (Figures 5, 7 and 8). Generally, land-cover change impact is greater in summer than in winter because changed vegetation characteristics (e.g. modified LAI, root depth, interception-storage capacity, stomatal resistance) affect evapotranspiration directly, while albedo and emissivity changes affect evapotranspiration indirectly via altered partitioning of net radiation. Owing to atmospheric water transport, local land-cover changes may have inter-regional consequences that increase when CO\(_2\) is doubled because of increased residence times. The potentially far-reaching impact of local land-cover changes in a warmer climate suggests that further investigations into the influence of land-cover on the climate system are urgently needed.

Our study shows that whether the assumed regional land-cover changes affect the NAO or PNA depends on

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Figure 11. As in Figure 9, but for the Ob region.

Figure 12. Vertical profile of domain-averaged relative humidity from the surface to 100 hPa for the Ob region in September. Solid and dashed lines represent control (CTR) and changed land-cover (LUC), respectively.

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Figure 12. Vertical profile of domain-averaged relative humidity from the surface to 100 hPa for the Ob region in September. Solid and dashed lines represent control (CTR) and changed land-cover (LUC), respectively.
INTERACTION OF DOUBLING CO₂ AND CHANGING REGIONAL LAND-COVER IMPACTS

Figure 13. As in Figure 9, but for the St. Lawrence region.

Figure 14. Differences between conditions of concurrently changed land-cover and doubled CO₂ (LUCDBL) - doubled CO₂ (DBL) in domain-averaged low-level CF (solid line), and long-wave radiation (LWR) loss at the top of the atmosphere (dashed line) for the St. Lawrence region in October. Note that CF ranges between 0 and 1. For convenience, the lines for zero changes in low-level CF and for LWR loss at the top of the atmosphere are shown as thick solid and thick dashed lines, respectively.

the season and region of altered land-cover. More in-depth research on the impacts of mid- or high-latitude land-cover changes on NAO or PNA is required to assess the global consequences of extra-topical land-cover changes.

Occasionally, significant non-linear interactions exist between the impacts of changing land-cover and doubling CO₂ on residence time. These significant interactions between the treatments result from the sensitivity of phase-transition processes to temperature, and hence sensitivity to temperature changes. Whether treatment interactions become significant for a given water-cycle-relevant quantity depend greatly on a region’s thermal, climatic, and hydrological regime, and on the season. Owing to the temporal and spatial heterogeneity of interaction between the impact of changing land-cover and doubling CO₂, conclusions drawn for the global scale can be misleading when applied to a region.

On the regional scale, during the cold season, the LUCDBL changes in runoff are broadly in phase with, and of similar magnitude to those of DBL for all regions (cf Figures 9, 11, 13 and 15). The same is true for cold season evapotranspiration increases in St. Lawrence, Ob, and Yukon. In spring, changes in LUCDBL and LUC residence times are in phase in the Yukon and Ob regions. In summer, changes in LUCDBL and LUC precipitation are in phase in the Ob and St. Lawrence regions. This similarity between LUC and LUCDBL means that local land-cover changes have comparable impact on the water cycles of these regions under reference and doubled CO₂ conditions.
In Colorado, land-cover changes significantly enhance sensible heat flux. Therefore, summer temperatures and atmospheric water demands increase more in LUCDBL than in DBL. At the same time, land-cover changes reduce evaporation of intercepted water and transpiration. Thus, evapotranspiration and precipitation decrease in LUCDBL compared to DBL. Consequently, we conclude that summer dryness will intensify water shortages in this semi-arid region if the assumed land-cover changes occur concurrently with doubling CO$_2$. Compared to the separate effects of changed land-cover and doubled CO$_2$ on residence time, their combined impact enhances June residence time in the St. Lawrence region. In the Yukon region, the combined treatment diminishes November evapotranspiration and September residence time. These results show that combining changed land-cover and doubled CO$_2$ may either diminish or enhance separate treatment impacts on water-cycle-relevant quantities. They also indicate that the effect of land-cover changes is not the same everywhere under altered CO$_2$ conditions. Obviously, whether the interaction between two treatments, doubling CO$_2$ and changing land-cover, is significant depends on complex thermal, dynamical, and hydrological preconditions.

How efficient a (feedback) mechanism can become in response to anthropogenic changes is also regionally dependent. For all four regions, the snow-albedo feedback, for instance, decreases winter and early spring near-SAT in areas of changed land-cover. Since the snow-albedo feedback depends heavily on the solar zenith angle, changed land-cover impact on snow-albedo feedback is limited by low high-latitude solar zenith angles during boreal winter. However, the effects of land-cover changes become apparent (1) during spring, when downward radiation increases while snow is still present and (2) where snow lasts longer than at mid-latitudes, or (3) in fall, when high-latitude snowfall occurs earlier than at mid-latitudes. Where land-cover has changed, reduced near-SATs result in more winter snowfall and increased spring runoff in LUCDBL compared to DBL. However, in the St. Lawrence region, for example, increased snow-albedo feedback resulting from land-cover changes lasted for a shorter time, due to both shorter duration of snow coverage than in high latitude regions or CTR, and lesser winter snowfall than in CTR.

Temperature increases due to doubling CO$_2$ may affect the partitioning of precipitation between liquid and solid forms. Consequently, following land-cover changes, the onset of the snow-albedo feedback effect occurs later in LUCDBL than in LUC. Thus, the impact of combined treatments depends on the delay of snow-cover onset (which relates to air temperature).

In the Ob and St. Lawrence regions, land-cover changes impact the spring runoff differently under reference and doubled CO$_2$ conditions. In the Yukon, land-cover changes hardly affect spring runoff under either
CO₂ condition, while in Colorado, both changing landcover and doubling CO₂ reduce the spring runoff. In regions governed by low-pressure systems external to the region (e.g., Yukon in winter), or where storm tracks advect large amounts of moisture into the region (St. Lawrence), doubling CO₂ largely increases low-level cloud-fraction and thereby reduces LWR loss. Consequently, the temperature impact of land-cover changes may not be noticeable if the two treatments occur concurrently. However, this interaction may not occur in semi-arid regions like Colorado, where the annually averaged low-level cloud-fraction is <10%.

Treatment interactions may also vary between water-cycle components. Evapotranspiration is more sensitive to land-cover changes and interactions between CO₂ concentration and land-cover changes than precipitation, because land-cover changes directly influence the evapotranspiration by altering surface wind speed, canopy interception-storage capacity, or transpiration. Nevertheless, despite the complexities of these interactions and the difficulties posed by the great variability that exists between regions in response to changing climate and landscape, the likely future increases in atmospheric CO₂ concentration and the vital importance of the global water cycle and regional water cycles to life on earth and within a region makes it clear that elucidating the characteristics of regional water cycles and accurately projecting their response to possible future climate and land-cover changes is a task of vital importance for the atmospheric science community. A sensitivity study performed with tripled CO₂ conditions without and with altered land-cover showed similar results, with the changes in the regional interaction between treatments and the slowing down of the regional water cycles being even further pronounced.

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