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Relationships between landscape, snowcover depletion, and regional weather and climate

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Abstract:

The effects of landscape changes on winter and spring snow-related processes, and on regional weather and climate are not thoroughly understood. In this study, a climate version of the Regional Atmospheric Modelling System (ClimRAMS) is used to investigate the effects of landscape change on seasonal snow depletion and its corresponding effects on atmospheric and hydrologic processes. Two simulations of the 1996 spring melt season in the Rocky Mountains and Northern Great Plains are compared. The first simulation utilizes the present-day vegetation distribution, and the second uses the same vegetation distribution with the exception that all forested regions are replaced by grassland. This vegetation modification affects 18% of the domain and changes the leaf area index, transmissivity of the vegetation canopy to incoming solar radiation, roughness length, and surface albedo. Additional snow-related differences occur because the snow lying over grass, and the snow under the forest canopy, exist in dramatically different radiative and thermal regimes. The snowcover changes resulting from the simulated deforestation influence the surface radiation balance, which leads to changes in surface sensible and latent energy fluxes, evaporation and transpiration rates, melt rates, and air temperature. The vegetation change also modifies snowcover depletion rates, which in turn cause variations in runoff production. Unmodified regions are affected through hydrologic transport processes. The manifestations of these changes with respect to regional weather and climate are discussed. Copyright © 1999 John Wiley & Sons, Ltd.

KEY WORDS deforestation; snowmelt; runoff; climate modelling

INTRODUCTION

The effects of landscape change have long been acknowledged as an important element of the climate system, and in recent years the climate-related effects of anthropogenic changes to the land surface have become an important research topic. Most researchers have concentrated on the global effects of tropical deforestation. The effects of landscape change on winter climates and hydrologic processes have not been thoroughly investigated.

The significance of specific landscape changes such as deforestation have been studied using global circulation models (GCM). These studies concentrate on two geographic areas, the tropics (Henderson-Sellers and Gornitz, 1984; Dickinson and Henderson-Sellers, 1988; Lean and Warrilow, 1989; Shukla *et al.*, 1990; Nobre *et al.*, 1991; Durbidge and Henderson-Sellers, 1993; Henderson-Sellers *et al.*, 1993), and the boreal forest (Thomas and Rowntree, 1992; Bonan *et al.*, 1995). Results from the tropical studies note a surface temperature increase and a precipitation decrease linked to tropical deforestation. In contrast, the results from boreal studies noted an air temperature decrease. Thomas and Rowntree (1992) also noted an increase in snow depth in deforested regions. Additionally, several of the authors have noted some type of

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teleconnection of the atmospheric fields from the deforested area to regions which remained unaltered (Dickinson and Henderson-Sellers, 1988; Durbidge and Henderson-Sellers, 1993; Henderson-Sellers *et al*., 1993; Copeland *et al*., 1996; Xue, 1996, Chase *et al*., 1996).

Some of these previous efforts have examined the regional effects of deforestation using global models (Lean and Warrilow, 1989; Henderson-Sellers *et al*., 1993; Bonan *et al*., 1995). However, the use of a regional climate model to examine the effects of landscape changes has been less common. Copeland *et al*. (1996) used a climate version of the Colorado State University, Regional Atmospheric Modelling System (ClimRAMS) to evaluate the effects of changing from the pre-settlement to the present-day vegetation distribution of the Continental United States. At that time, ClimRAMS was not able to adequately deal with snowcover or snowmelt, therefore their study concentrated on July 1989. The vegetation changes imposed by that study altered the surface albedo and thus the net radiation available at the surface. This resulted in an increase in mean daily temperature for the case of present-day vegetation. In addition, they linked changes in leaf area index, fractional vegetation coverage, and the balance between sensible and latent energy fluxes to whether individual areas warmed or cooled. Copeland *et al*. (1996) also noted an increase in precipitation throughout the domain when the pre-settlement vegetation was replaced by the present-day vegetation distribution.

In all of the above-mentioned studies, the authors first completed a control simulation using the best available representation of vegetation and atmospheric conditions. Next, they compared this control simulation to a second simulation which was identical except for a change in the vegetation distribution. The study presented herein makes use of this well-established methodology.

Recent improvements to ClimRAMS allow the model to successfully simulate both winter and summer climates (Liston *et al*., 1998). The addition of snowcover parameterizations and a snowmelt scheme have greatly improved the model's representation of the winter to spring seasonal transition. This study takes advantage of these advances in order to analyze the snow-related effect of forestation/deforestation in the Rocky Mountains.

In this study, we examine the depletion of seasonal snowcover as an initial value problem. Liston *et al*. (1998) concludes that models run at coarse resolution (e.g. grids with a 50 km Δx) are unable to adequately simulate snow accumulation processes in complex terrain because the processes which drive that accumulation occur at smaller spatial scales than represented by the model. Therefore, during this investigation we use the observed snow distribution on 26 January 1996, for the model initial condition. The model continues to accumulate snow in some regions of the domain for the remainder of the winter (approximately six weeks). Since the emphasis of this study is to examine the snowcover depletion, the late January start date has been chosen to minimize the influence of snow-accumulation deficiencies, while still including all the ablation period.

RAMS SUMMARY

RAMS was developed at Colorado State University to facilitate research into predominantly mesoscale and regional, cloud and land-surface atmospheric phenomena and interactions (Pielke, 1974; Tripoli and Cotton, 1982; Tremback *et al*., 1985; Pielke *et al*., 1992; Walko *et al*., 1995a). This model is fully three-dimensional; nonhydrostatic (Tripoli and Cotton, 1980); includes telescoping, interactive nested grid capabilities (Clark and Farley, 1984; Walko *et al*., 1995b); supports various turbulence closure (Deardorf, 1980; McNider and Pielke, 1981; Tripoli and Cotton, 1986), short and long wave radiation (Mahrer and Pielke, 1977; Chen and Cotton, 1983, 1987; Harrington, 1997), initialization (Tremback, 1990), and boundary condition schemes (see Pielke *et al*., 1992); includes a land-surface energy balance submodel which accounts for vegetation-, open water-, and snow-related surface fluxes (Mahrer and Pielke, 1977; McCumber and Pielke, 1981; Tremback and Kessler, 1985; Avissar and Mahrer, 1988; Lee, 1992; Liston *et al*., 1999); and includes explicit cloud microphysical submodels describing liquid and ice processes related to clouds and precipitation (Meyers *et al*., 1992; Meyers, 1995; Schultz, 1995; Walko *et al*., 1995a). In numerous publications RAMS has been shown to successfully simulate weather processes.

In order to successfully simulate winter and spring climates, the model must account for the effects of seasonal snowcover. Recent additions to ClimRAMS have made this possible (Liston *et al.*, 1999). In the model, snow is accumulated on the ground when the temperature of the lowest layer in the atmospheric model is less than 0 °C. The snowcover is represented by one layer of constant density and thermal properties. The snowcover modifies the surface albedo depending on whether the snow is dry (albedo = 0·8) or melting (albedo = 0·6) (Cline, 1997; Gray and Male, 1981); for thin snowcovers (less than 5 cm snow-water-equivalent) the albedo decreases linearly with decreasing snow depth. The presence of snow modifies the ground heat flux computation, and the surface roughness. The energy available to melt snow is computed as part of the surface energy balance. When the snow melts it contributes directly to increasing the soil moisture. Snow is assumed to fall through the vegetative canopy where it modifies the under-canopy radiation budget; interception by the vegetative canopy is not taken into consideration. Each vegetation type is assigned a characteristic height, and once the snow depth exceeds that height, the surface characteristics become that of snow rather than the vegetation.

The ClimRAMS land-surface scheme includes all the terms required to satisfy a surface energy balance. Incoming shortwave radiation is a function of time of day and year, and is attenuated as it passes through the atmosphere. Longwave radiation is emitted and absorbed by the atmosphere. The vegetative canopy reduces the amount of radiation incident on the surface. The longwave flux from the surface is a function of surface temperature. The latent energy flux is a function of atmospheric pressure, air density, wind speed, and the vapor pressure gradient above the surface. The sensible energy flux is a function of air temperature, stability, wind speed, and the temperature gradient between the lowest level of the atmosphere and the surface. The surface energy balance is solved to yield the surface temperature. In the presence of snow, surface temperatures greater than 0 °C resulting from the surface-energy balance indicates that energy is available for melting. The amount available is then computed by setting the surface temperature to 0 °C and recomputing the surface-energy balance.

EXPERIMENTAL DESIGN

This study uses ClimRAMS to simulate the evolution of the snow and atmosphere from 26 January, 1996, through 30 June, 1996, a period of 157 days. For these simulations the model uses two nested grids. Grid 1 covers the Continental United States at a horizontal grid spacing of 200 km ($\Delta x = \Delta y = 200$ km) (Figure 1). Grid 2 is nested within Grid 1 and covers the Central Rocky Mountains and the Northern Great Plains at a horizontal grid spacing of 50 km ($\Delta x = \Delta y = 50$ km) (Figure 1). The model contains 20 vertical levels. The

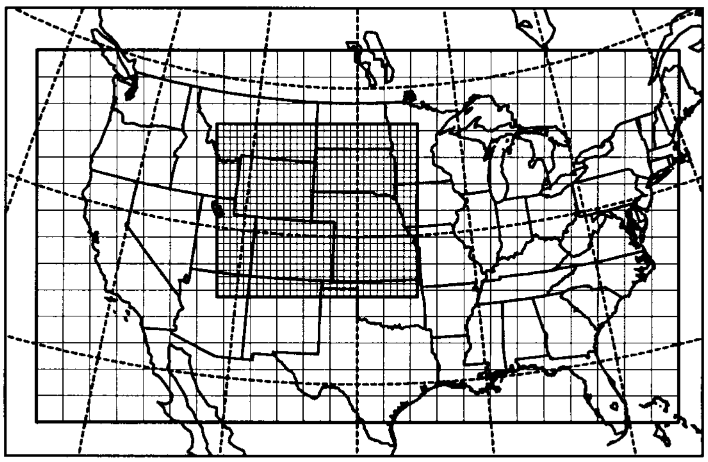

Figure 1. RAMS grids used in simulation, Grid 1 $\Delta X = \Delta Y = 200$ km, Grid 2 $\Delta X = \Delta Y = 50$ km

spacing of the vertical levels was 0·250 km at the lowest level and increases with increasing altitude but never exceeds 2 km. Model integrations were performed using a time step of 120 seconds. The particular grid configuration and model setup used in the current study has produced results which compare well with observed atmospheric fields (Liston *et al*., 1999).

The model uses global, six hourly, reanalysis data sets from the National Centers for Environmental Prediction (NCEP) (Kalnay *et al*., 1996) to define the lateral boundary conditions. The reanalysis data are available on a 2·5° latitude by 2·5° longitude vertical pressure-level grid. These data are interpolated to Grid 1, and used to update the lateral boundary conditions of this coarsest grid every six hours. This technique forces the large scale atmospheric patterns in the model to be consistent with observed patterns, while allowing the model to adjust to surface changes within the domain and alter small scale processes. Thus we are able to examine alteration in the melt cycle and surface energy balance while maintaining the observed large scale atmospheric forcing. Atmospheric initial conditions are also provided by the NCEP reanalysis data sets. Since the land-surface has been altered for these simulations (e.g. snowcover initial conditions, deforestation) the planetary boundary layer needs to adjust to the changes in the underlying surface. The adjustment time scale is on the order of hours, and hence takes place early during our simulations.

The snow-water-equivalent (swe) initial conditions, provided on 26 January, 1996 (Figure 2) are a compilation of two data sets: swe distributions from the National Operational Hydrologic Remote Sensing Center (NOHRSC) for the Western United States (approximately west of 100°W Longitude), and snow depths from the National Climate Data Center (NCDC) Summary-of-the-Day data set for the remainder of the domain. The NCDC snow depths were converted to swe by defining representative snow densities to each

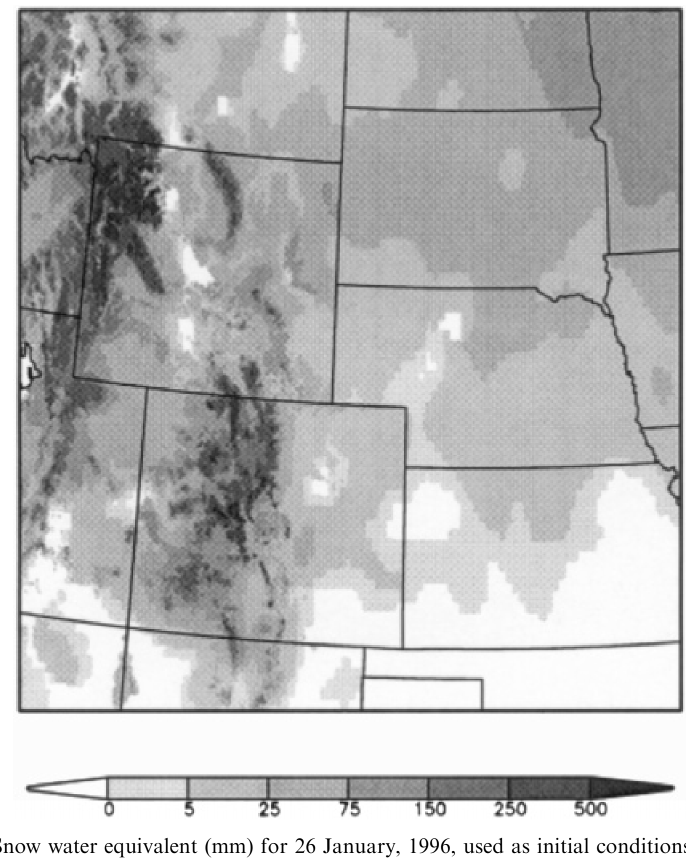

Figure 2. Snow water equivalent (mm) for 26 January, 1996, used as initial conditions for Grid 2

Table I. Snow densities applied to the Sturm *et al.* (1995) snow classifications and used in conversion from snow depth to snow water equivalent

Classification	Density (kg/m^3)
Tundra	280
Taiga	225
Maritime	300
Ephemeral	350
Prairie	250
Alpine	275

of the snow classes mapped by Sturm *et al.* (1995). The densities assigned to each snow class are given in Table I.

All of the snow distribution data comes from observed data sets. Since there has been no large scale deforestation in the Central Rocky Mountains, the snow distribution data used in the model predominantly accumulated in forested regions. Thus we can make no assessment as to how snow accumulation processes differ due to deforestation. Our intent is to examine the alteration in terms of surface energy and snowmelt cycles. In areas of complex terrain it is very difficult to accurately map the distribution of snow-water-equivalent. Although the initial snow-water-equivalent conditions are certainly not absolutely correct, in compiling this data set we made every attempt to use the most accurate information available at the scales of interest.

The model also requires an initial soil moisture distribution. Spatial distribution of soil moisture in complex terrain is also difficult to know quantitatively. Liston *et al.* (1999) used an identical grid configuration to successfully simulate the 1996 winter season. For these simulations the model soil moisture field was initialized with the same field used by Liston *et al.* (1999).

The model used the present-day vegetation distribution, and the standard RAMS methodology to define the vegetation type, areal fraction and leaf area index (Liston *et al.*, 1999). The vegetation-classification data were obtained from the EROS Data Center Distributed Archive Center (EDC DAAC), located at the US Geological Survey's EROS Data Center in Sioux Falls, South Dakota (http://edcwww.cr.usgs.gov/landdaac/).

Two separate simulations were completed for this study. The first simulation (Run 1) utilized the present-day vegetation distribution (EDC DAAC). The second simulation (Run 2) utilized the same vegetation data set, however areas covered with evergreen trees, deciduous trees, and mixed woodlands (collectively referred to as trees) were changed to short grass prairie (referred to as grassland) (Figure 3). The vegetation change affected 18% of the Grid 2 domain. In the model, changing the vegetation type alters the surface representation by modifying the following parameters: albedo, roughness, transmissivity of the vegetation canopy to incoming solar radiation, and leaf area index.

RESULTS

Changing the natural vegetation distribution has a large effect on the surface energy budget. The most-salient consequence of the landscape change is a change in surface albedo (Figure 4). Under snow-free conditions changing trees to grassland has the effect of increasing the surface albedo (since the grassland albedo is large compared to that of trees (Table II)). The difference in albedo, between grasslands and trees, is even more dramatic when snow is present. For grassland the vegetation is snow covered, and the albedo for those areas is equal to that of snow. For the wooded areas, the snow is masked by the protruding vegetation, and the albedo seen by the atmosphere equals that of trees. The contrast in surface characteristics from snow to forest is significant enough to induce mesoscale circulations (Taylor *et al.*, 1998). In RAMS, for each vegetation type, there is also a vegetation-fraction parameter, which defines the fraction of each grid cell

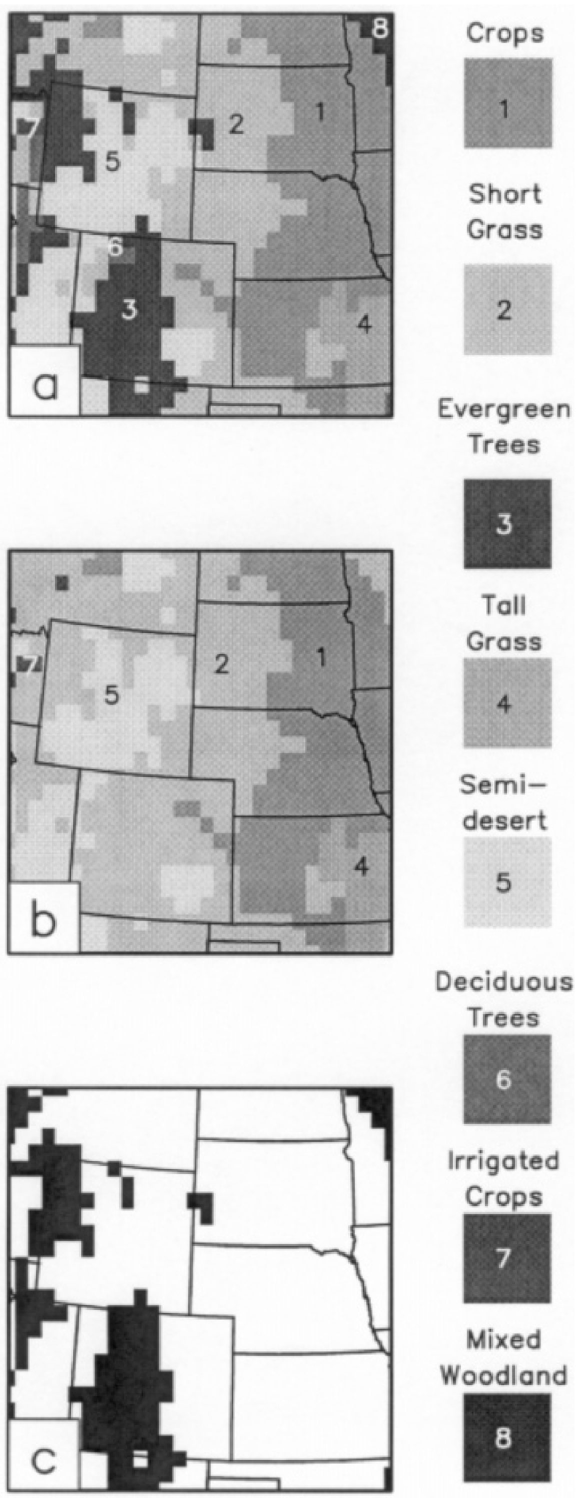

Figure 3. Vegetation type and distribution for simulations, (a) Present-day vegetation used in Run 1, (b) Modified vegetation used in Run 2 (trees replaced by short grass), (c) 18% of the domain was affected by the vegetation modification

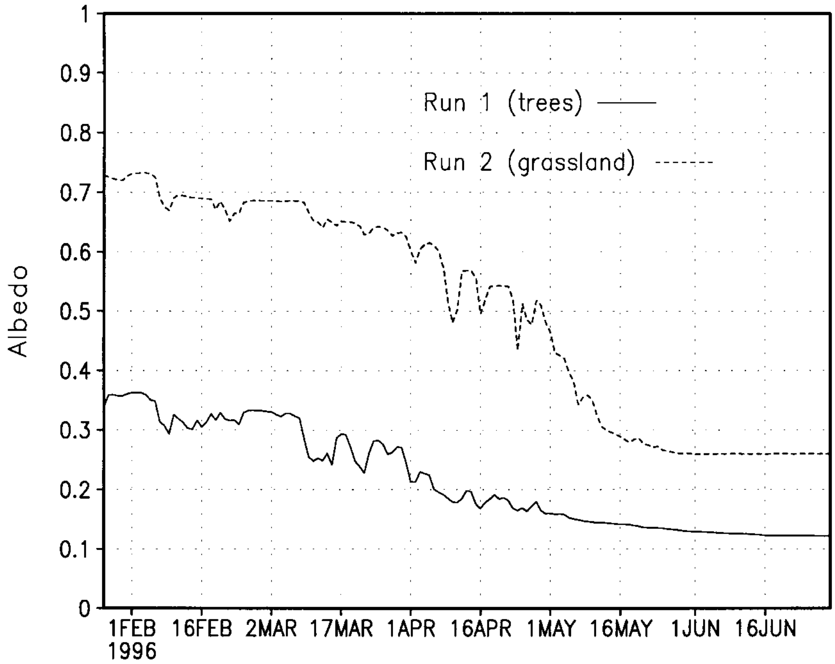

Figure 4. Area-averaged albedo over the modified areas

Table II. Model vegetation parameterization values

Vegetation type	z_0	α	Max LAI
Crop/mixed farming	0·06	0·20	6
Short grass prairie	0·02	0·26	2
Evergreen needleleaf tree	1·00	0·10	6
Deciduous needleleaf tree	1·00	0·10	6
Deciduous broadleaf tree	0·80	0·20	6
Evergreen broadleaf tree	2·00	0·15	6
Tall grass prairie	0·10	0·16	6
Desert	0·05	0·30	0
Tundra	0·04	0·20	6
Irrigated crop	0·06	0·18	6
Semidesert	0·10	0·25	6
Ice cap/glacier	0·01	0·40	0
Bog or marsh	0·03	0·12	6
Inland water	0·0024	0·14	0
Ocean	0·0024	0·14	0
Evergreen shrub	0·10	0·10	6
Deciduous shrub	0·10	0·20	6
Mixed woodland	0·80	0·18	6
Bare ground	0·05	0·14–0·31*	

z_0 = roughness length (m), α = albedo, LAI = leaf area index.
*The albedo of bare ground depends on soil moisture levels.
Note: Vegetation types not represented in Figure 2 were used in Grid 1.

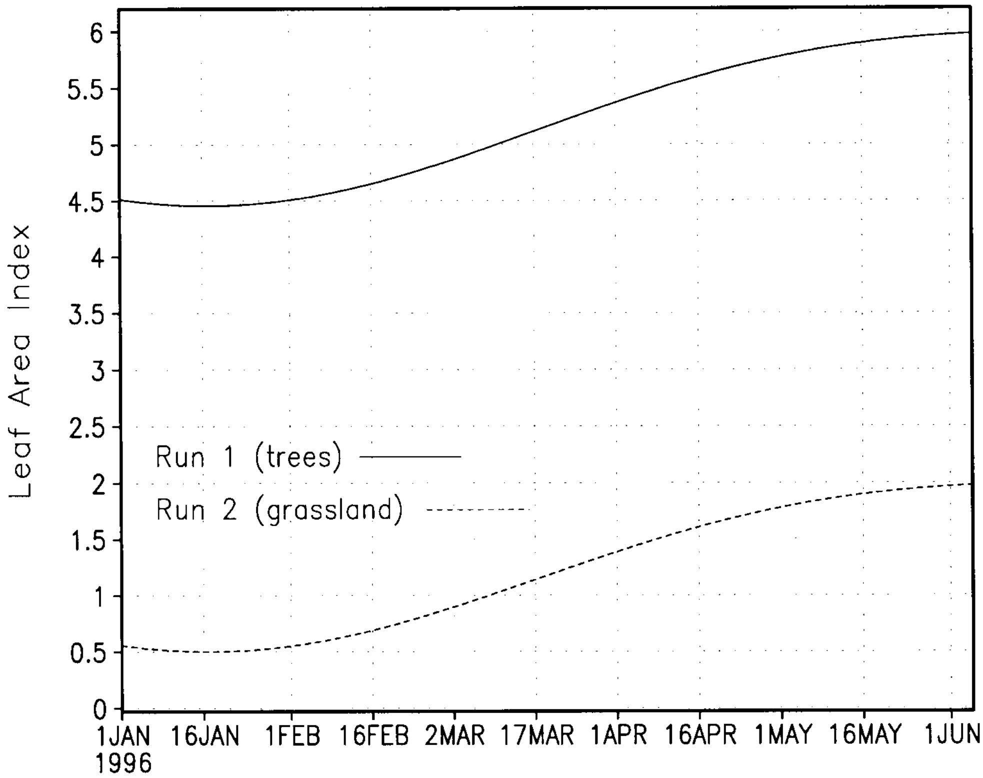

Figure 5. Area-averaged leaf area index over the modified areas

covered by vegetation. The remaining fraction is assumed to have no vegetation. Thus any snow in this portion of the grid cell is precluded from being masked by protruding vegetation. As the spring melt season progresses, the albedo gradually decreases until its value is only due to the fractional weighing of the vegetation and bare ground (Figure 4). Altering the vegetation type also changes the transmissivity of the vegetative canopy. For the alterations made to Run 2 (deforestation), this has the effect of reducing the net radiation. The deforestation also altered the partitioning in the surface energy terms.

The leaf area index for trees is higher than that of grasslands (Table II), and as the growing season progresses, the leaf area index increases from a winter minimum to a summer maximum (Figure 5). The differences in leaf area index affects atmosphere-landscape interactions by altering both evaporation and transpiration processes. In the model and in the real-world evaporation and transpiration are two separate processes, however we have not examined them independently, and henceforth will refer to them collectively as evapotranspiration.

In the absence of trees, the leaf area index of the region decreased (Figure 5). Thus, the evapotranspiration rates for the altered areas are much greater in Run 1 due to the differences in leaf area index (Figure 6). The amount of evapotranspiration increases in the spring and early summer due to the increased leaf area index. The evapotranspiration values in Run 1 are always greater than those in Run 2.

The surface energy budget is also affected by changes in albedo and leaf area index. These effects are examined by looking at the surface latent and sensible energy fluxes. Comparing the areas modified by the deforestation, the sensible energy flux was greater in Run 1 than Run 2. In the absence of trees, the snow covers the vegetation. As a result the surface albedo is that of snow rather than that of the land surface or vegetation (Figure 4). Again comparing the modified regions, the latent energy flux in Run 1 is higher during the entire simulation (Figure 6). During the winter, differences in latent energy flux between Runs 1 and 2

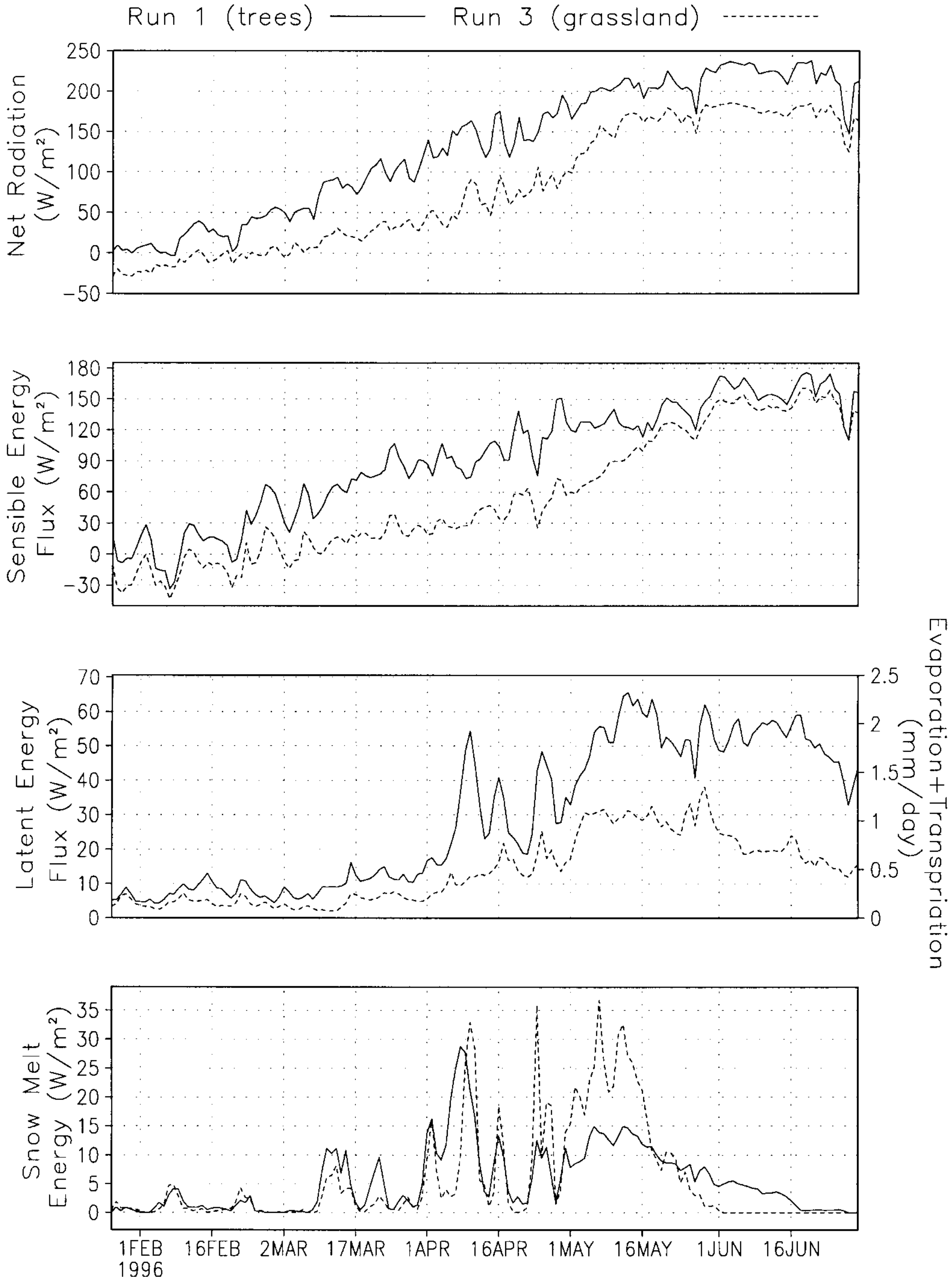

Figure 6. Area-averaged values of all terms, over the modified areas, in surface energy balance computation (conductive flux not shown)

over the modified areas are small. In spring, with the increased leaf area index in Run 1, the latent energy flux differences become much larger.

The air temperature in the lowest portion of the atmosphere is dependent on the surface energy fluxes. Although the sensible heat flux is always greater in Run 1 (Figure 6), there is not a constant difference in

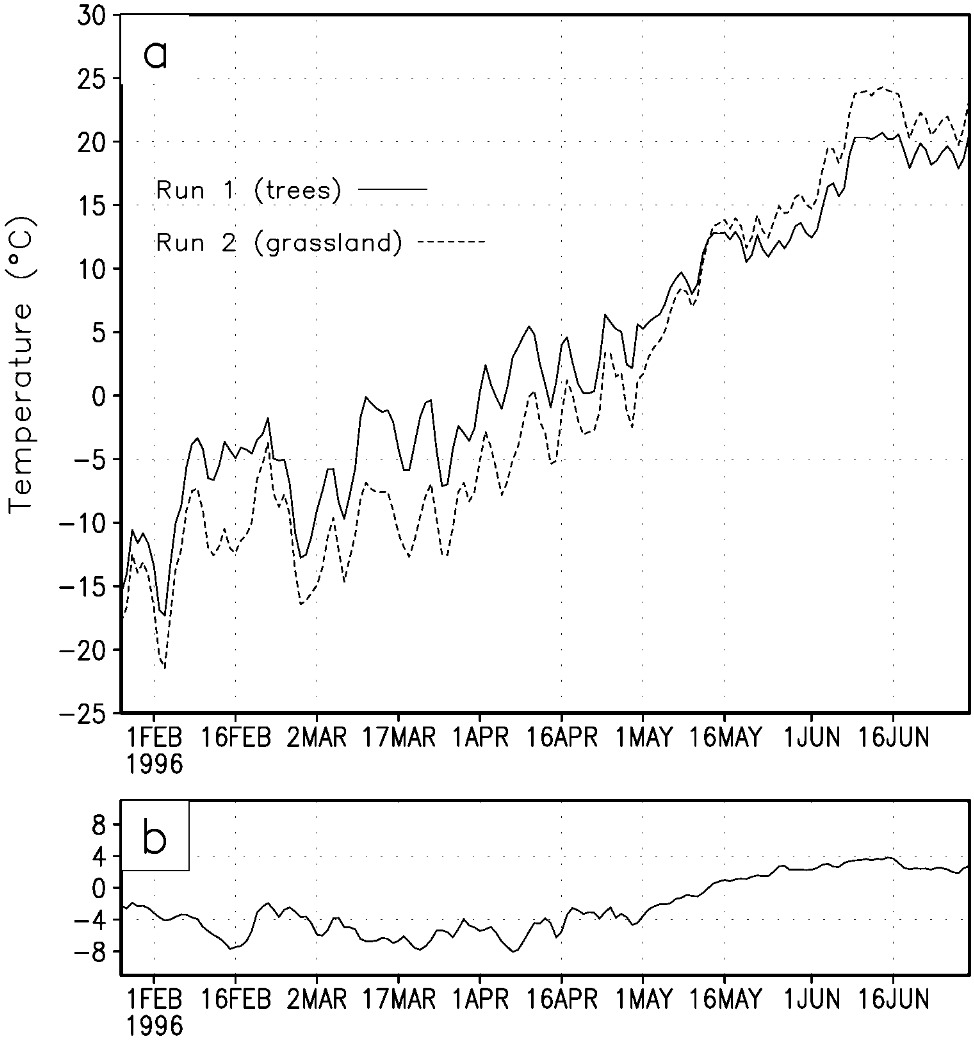

Figure 7. (a) Area-averaged screen height (2 m) air temperature over the modified areas, (b) Difference in area-averaged air temperature, Run 2 – Run 1

surface air temperature between the two simulations (Figure 7). During the winter, the air temperatures are higher in Run 1 over the areas which were modified for Run 2. The lower albedo of the forested surface results in greater sensible energy fluxes. The greater sensible energy fluxes contribute to higher surface air temperatures in the forested areas during winter. In late May, the surface air temperatures in Run 2 exceed those in Run 1. This transition occurs during the melt period. Once the snow is removed and the melt energy decreases (Figure 6) there is more energy partitioned into latent and sensible energy flux. This increase contributes to increased temperatures in Run 2 during the end of May. As the snow melts the underlying surface is exposed, and the albedo of the surface also changes accordingly. The seasonal change from winter to summer affects the albedo of both simulations due to the snow melting. However, the absolute value of the change is larger for the grassland areas (Run 2), and even though the areas changed to grassland (modified areas in Run 2) always have a higher albedo, the decrease in albedo and increase in sensible energy flux contribute to higher surface air temperatures during the spring and summer.

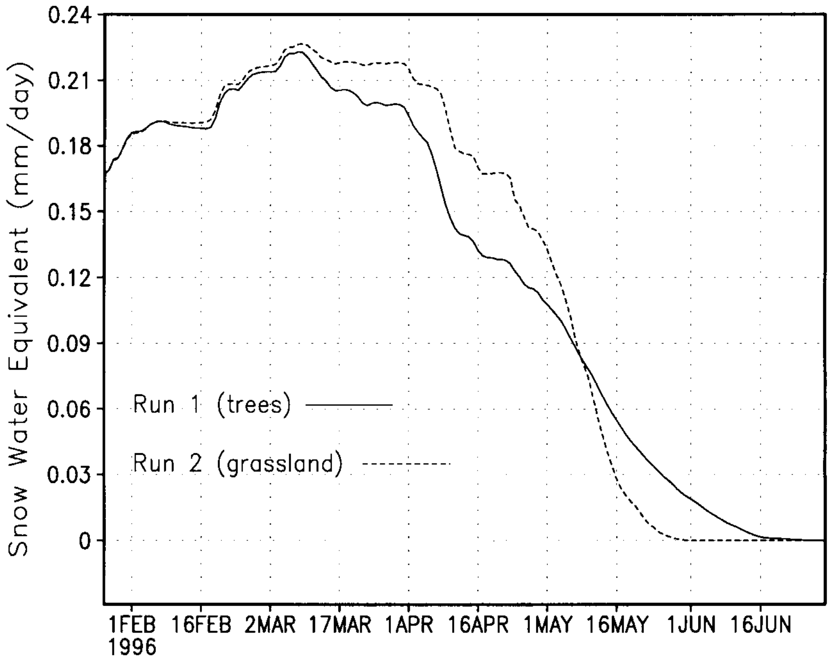

Figure 8. Area-averaged snow water equivalent over the modified areas. The snow-free date for Run 1 was 27 June, and for Run 2 it was 1 June

In these simulations, there is not a large difference in precipitation amounts or distribution (Table III). This is mainly due to the time of year which the simulations represent. During the winter, precipitation mechanisms are dominated by synoptic-scale circulations. Therefore, the precipitation patterns are mostly controlled by large-scale forcing, and not greatly affected by local changes in surface properties. During the summer, in the middle-latitudes, continental precipitation is largely due to convective circulations. These circulations are much smaller in scale and are affected more by local vegetation changes than those occurring during winter (Lee, 1992). Summer-time simulations, and studies based in the tropics, have shown that landscape alterations can definitely affect precipitation values and distributions (Henderson-Sellers *et al.*, 1993; Copeland *et al.*, 1996).

The landscape alterations in these simulations have a dramatic effect on the rate of snow depletion (Figure 8). During January and February, the model accumulates snow at nearly the same rate over both the grasslands and wooded areas. As the spring-melt cycle is initiated, the melt rate is greater in the wooded areas. However, as the melt cycle progresses the grassland melt rate increases, surpassing that of the wooded areas by 1 May. The snow melts off rapidly in the grasslands, and by 1 June the snowcover has completely melted. The snow melted during a much shorter period over the areas where the landscape changes were imposed, and this redistribution of melt energy had a dramatic effect on the snow-free date. The snow-free date in Run 1 was 27 June, while in Run 2 the snow-free date occurs on 1 June, nearly four weeks earlier.

Alterations in the seasonal melt rates, due to landscape changes, also modify spring runoff (Figure 9). In Run 1 the runoff occurs over a much longer time period, while in Run 2 the peak runoff is substantially increased. Runoff values for the two simulations are near equal until the second week in March. Near the end of April the runoff values for Run 2 become much greater than the corresponding values from Run 1. During May, the majority of snow over the grassland areas melts and runs off. By June 1 the spring runoff in Run 2 is reduced to a trickle. However, in Run 1 the spring runoff continues until the end of the simulation.

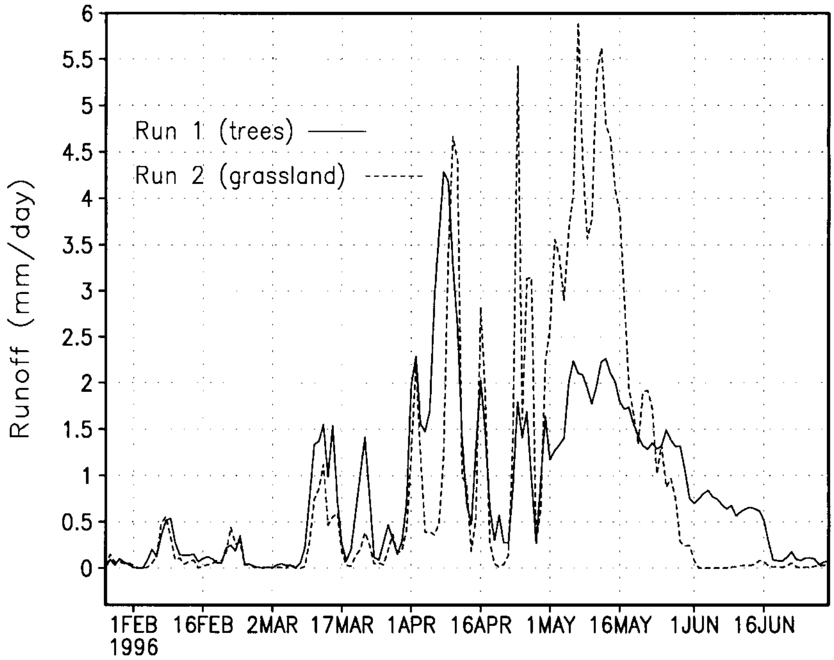

Figure 9. Area-averaged runoff production over modified areas

DISCUSSION AND CONCLUSIONS

Interactions between the atmosphere and landscape can play an important role in the regional climate system. Changes in vegetation type alter these interactions and affect a wide range of atmospheric and hydrologic fields. The effects of landscape change are not limited to a change in vegetation. The new vegetation distribution leads to changes in atmospheric temperature and moisture fields, and alter hydrologic distributions of soil moisture and runoff magnitude and timing.

For this winter simulation, a change in the vegetation type from forest to grassland had several effects. The albedo of the surface was affected by the vegetation type, and whether the snow lies over or under the vegetation. The net radiation available at the surface was also altered by the change in vegetation type. This had a large impact on the surface energy budget and other atmospheric fields. The leaf area index was very different for trees and grasses and, as the growing season evolved, this difference increased, leading to significant differences in evapotranspiration. The surface roughness was also dependent on the vegetation type, impacting the turbulent transport of energy in the atmospheric wind fields (not shown). These local changes in surface properties impacted both atmospheric and hydrologic processes. The temperatures in the deforested simulation (Run 2) were lower in the winter and higher in the summer. Also, the moisture returned to the atmosphere decreased from the forested simulation (Run 1) by 41%, and the peak runoff increased by 27%. As a result of the deforestation, the runoff season decreased by nearly four weeks.

Since the timing and magnitude of runoff production is altered by deforestation (e.g. Run 2), moisture and energy are transferred to adjacent regions (grid cells), via river and stream networks, at different rates than would have occurred without the deforestation (e.g. Run 1). The model calculates a runoff value for each grid cell based on surface energy and moisture balances, but unfortunately the model is not equipped with a runoff routing scheme. Implementation of such a scheme would allow closer evaluation of the impact on downstream areas. These simulations show that if large-scale deforestation ever occurred it could have significant implications regarding water-use policy. As an example, increased runoff during a shorter time period would impact reservoir storage, flood control and irrigation management plans.

Deforestation has also been shown to have an effect on snow accumulation. Kattelman (1990), Berris and Harr (1997), and Harr (1986) have investigated the effects of clear cutting on snow accumulation and melt cycles. Although this effort did not attempt to examine the effects of deforestation on snow accumulation, the results from this study agree with the melt cycle trends these previous studies observed. These researchers found that snow accumulation increased in clear-cut areas. This occurs in part from interception by vegetation canopies and the snow-related processes that occur within and under the vegetation canopy. Although the model does account for radiative interactions within and under the vegetation canopy, the model used in this study does not simulate the snow-related process that occur within the vegetation canopy (e.g. snow interception). Accounting for such processes would allow the model to more accurately simulate physical differences, due to deforestation, for both the accumulation and melt periods.

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