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Sensitivity of June near-surface temperatures and precipitation in the eastern United States to historical land cover changes since European settlement

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[1] Land cover changes alter the near surface weather and climate. Changes in land surface properties such as albedo, roughness length, stomatal resistance, and leaf area index alter the surface energy balance, leading to differences in near surface temperatures. This study utilized a newly developed land cover data set for the eastern United States to examine the influence of historical land cover change on June temperatures and precipitation. The new data set contains representations of the land cover and associated biophysical parameters for 1650, 1850, 1920, and 1992, capturing the clearing of the forest and the expansion of agriculture over the eastern United States from 1650 to the early twentieth century and the subsequent forest regrowth. The data set also includes the inferred distribution of potentially water-saturated soils at each time slice for use in the sensitivity tests. The Regional Atmospheric Modeling System, equipped with the Land Ecosystem-Atmosphere Feedback (LEAF-2) land surface parameterization, was used to simulate the weather of June 1996 using the 1992, 1920, 1850, and 1650 land cover representations. The results suggest that changes in surface roughness and stomatal resistance have caused present-day maximum and minimum temperatures in the eastern United States to warm by about 0.3°C and 0.4°C, respectively, when compared to values in 1650. In contrast, the maximum temperatures have remained about the same, while the minimums have cooled by about 0.1°C when compared to 1920. Little change in precipitation was found.

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1. Introduction

[2] Turbulent sensible and latent heat fluxes and longwave radiation generated from absorption of solar radiation by the land surface are the source of much of the energy that drives Earth's weather and climate. The magnitudes of these fluxes over the continents are strongly dependent upon the characteristics of the land cover, such as albedo, Leaf Area Index (LAI), stomatal resistance, roughness length, and vegetation fraction. Surface albedo controls the amount of solar radiation absorbed at the surface and therefore strongly modulates the amount of energy available for conversion to sensible and latent heat fluxes and longwave radiation. Darker vegetation has a lower albedo and absorbs more solar radiation than lighter vegetation, increasing the amount energy available for transfer to the lower atmosphere. Increasing the LAI can increase the latent heat flux by enhancing transpiration, as long as sufficient water is available in the root zone. Vegetation with lower stomatal resistance transpires greater amounts of water than plants with higher values of stomatal resistance when this water is present. Turbulent sensible and latent heat fluxes are also dependent upon the aerodynamic roughness length. Rougher surfaces enhance turbulent eddy formation and thus increase the energy transferred by turbulent sensible and latent heat fluxes. Finally, the vegetation fraction determines the amount of incoming radiation reaching the soil. The vegetation intercepts more of the incoming solar radiation when the vegetation fraction is higher. Because soil and vegetation have much different heat capacities this has a significant influence on near-surface temperature.

[3] There have been many studies in recent years illustrating how historical changes in land cover can alter weather and climate as summarized in the work of *Pitman* [2003], *Kabat et al.* [2004], *National Research Council* [2005], and *Cotton and Pielke* [2007]. In the study of *Copeland et al.* [1996], for example, the July 1989 weather was simulated for the present-day and presettlement land cover using CLIMRAMS. They found that average daily

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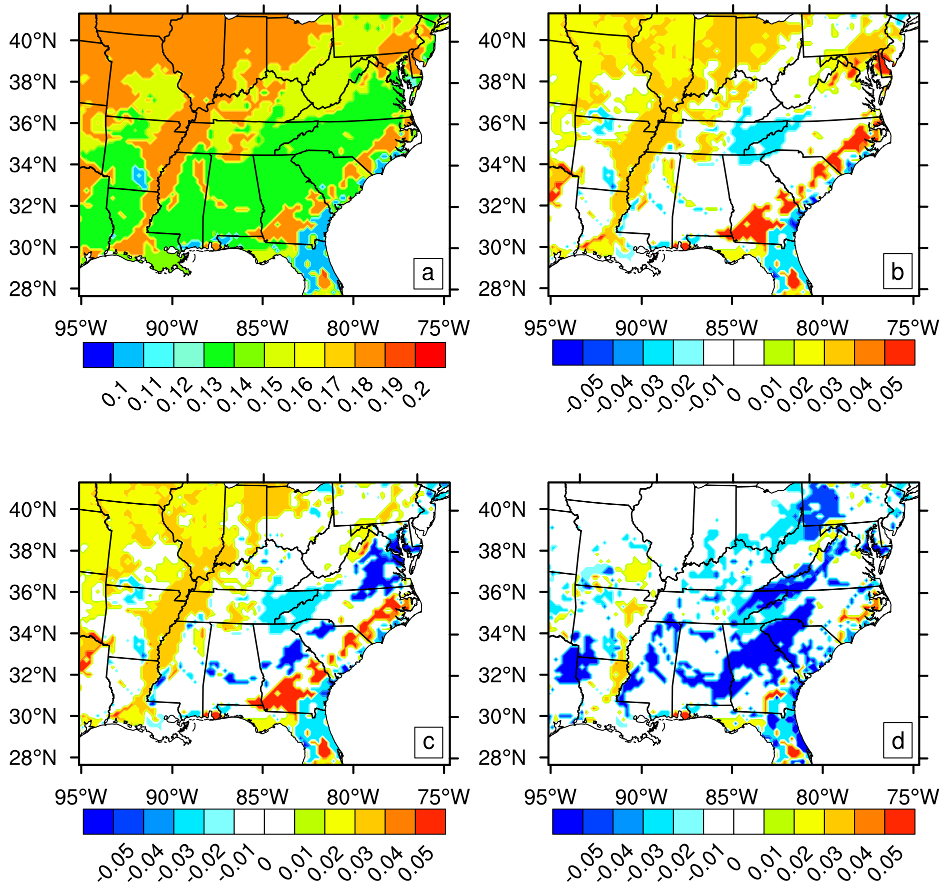

Figure 1. (a) The 1992 land cover broadband shortwave albedo. Also shown are differences in albedo between 1992 and (b) 1650, (c) 1850, and (d) 1920. Derived from land cover and biophysical parameters [*Steyaert and Knox*, 2008].

temperatures warmed by 0.05 K across the continental United States. They used the United States Geological Survey (USGS) 1990 1-km landcover database [*Loveland et al.*, 1990] for the current land cover and cover type aggregated from the *Küchler* [1995] potential natural vegetation data set for the pre-European settlement case.

[4] *Bonan* [1997] used the National Center for Atmospheric Research (NCAR) Land Surface Model (LSM v. 1.0) coupled with a modified version of the NCAR Community Climate Model version 2 (CCM2) to examine the sensitivity of the annual climate in the United States to changes in vegetation since European settlement. The modern vegetation was derived from the *Olsen et al.* [1983] data set and the natural vegetation was derived from a map of Küchler's potential natural vegetation [*Espenshade and Morrison*, 1990]. Using this global model he found that mean summer temperatures cooled by up to 1.5°C over parts of the eastern United States.

[5] *Baidya Roy et al.* [2003] used the Regional Atmospheric Modeling System (RAMS) equipped with the Land Ecosystem-Atmosphere Feedback model version 2 (LEAF-2) to simulate changes in United States summer weather owing to land cover change. They used land cover derived form the Ecosystem Demography (ED) model for 1700, 1910 and 1990 [*Hurtt et al.*, 2002]. The weather for the month of July from six years (1990–1995) was simulated. They found that the mean July temperature decreased slightly, generally less than 0.3°C, over most of the eastern United States between 1910 and 1990. In contrast when compared with 1700 the 1990 mean July temperatures were slightly warmer, mostly less than 0.6°C or less, over the eastern United States.

[6] An enhanced reconstructed historical land cover and biophysical parameter data set for the eastern United States [*Steyaert and Knox*, 2008] provides an excellent opportunity to further investigate the effects of long-term historical land cover change on weather and climate. *Steyaert and Knox* [2008] included a detailed discussion of methods used to develop this new data set. Basically, they used a mutually consistent set of 36 land cover classes and associated biophysical parameters to characterize the diversity of land cover conditions across the eastern half of the United States (i.e., land area to the east of 97° W) at the 1650, 1850, 1920, and 1992 time slices [see *Steyaert and Knox*, 2008, Tables 1 and 2]. The land cover for each time slice is determined by a set of fractional area land cover layers that sum to 1.0 at each location. Specifically, there are 22 fractional area land cover maps for 1650, 30 for 1850, 29 for 1920, and 26 for 1992 [see *Steyaert and Knox*, 2008, Table 1]. In addition to mapping the land cover condition associated with major land use transformations at these four time slices within the eastern United States, this new data set was designed to be

an improvement over existing data sets by using an expanded set of land cover and biophysical parameter classes that more fully account for regional differences in land cover characteristics owing to such factors as climate, topography, and soils [*Steyaert and Knox*, 2008].

[7] On the basis of the results of a land use intensity analysis, *Steyaert and Knox* [2008] described the spatial patterns of land cover condition including changes in time associated with the 1650, 1850, 1920, and 1992 time slices. The 1650 time slice was selected to depict the natural vegetation at a time when the influences of Native Americans or Europeans represented a recent relative minimum [*Steyaert and Knox*, 2008]. Their results showed that approximately 70% of the 1650 vegetation was relatively unchanged in 1850. Elsewhere clearing of the forest for wood products and agricultural farmland led to regenerating forest, mixed agriculture, village, and city land use patterns concentrated along the eastern seaboard and in the Ohio River basin. They showed that by 1920 approximately 90% of the eastern United States was transformed by intensive land use that led to a young regenerating forest, disturbed land with sparse vegetation, mixed agriculture, and growing city, residential, and urban areas. By 1992, the eastern forest had substantially regenerated including regrowth on abandoned farmland. Agricultural production had relocated according to land use suitability such as to lands supporting intense mechanized farming in the lower Mississippi River Valley and the corn-soybean belt states of the Upper Midwest. The landscape also became more fragmented with the growth and expansion of residential and urban development.

[8] *Steyaert and Knox* [2008] built on the LEAF-2 land data [*Walko et al.*, 2000] to develop an enhanced biophysical parameter table that contains a set of 10 biophysical parameters for each of the 36 land cover classes. These parameters include broadband solar albedo, land surface emissivity, LAI, seasonal change in LAI, maximum vegetation cover, seasonal change in maximum vegetation cover, aerodynamic roughness length, zero-plane displacement height, vegetation rooting zone depth, and canopy height. Each land cover class description is uniquely defined by its associated set of biophysical parameters (for parameter values, see *Steyaert and Knox* [2008, Table 2]). The fractional area land cover layers and biophysical parameter table were designed for easy incorporation into soil-vegetation-atmosphere-transfer models (SVATS). Figures 1 and 2 were derived from the *Steyaert and Knox* [2008] data set and show how the broadband solar albedo and roughness length have changed since 1650. Figure 1 shows that the albedo increased with the replacement of old growth forest by crops between 1650 and 1920. By 1992 the albedo had decreased again, although not to 1650 levels, as the forest regenerated. Figure 2 shows that the surface roughness length was greatly reduced from 1650 to 1920 owing to the clearing of the forest, however a small recovery has taken place since 1920 as eastern croplands were abandoned. *Steyaert and Knox* [2008, section 3.1 and Figures 7–11] provide a more detailed discussion of the regional patterns of land cover condition including temporal changes, land use intensity maps, and biophysical parameter maps for each time slice.

[9] Artificial drainage of wetlands since the 1700s was an important land transformation within the eastern United States, with one estimate of the fractional coverage decreasing from approximately 20% to 8% between the 1780s and 1980s [*Steyaert and Knox*, 2008]. The wetlands information in historical data sets such as the Küchler's potential natural vegetation [*Küchler*, 1964] has limitations, and the characterization and mapping of present-day wetlands is challenging and an ongoing research activity [*Steyaert and Knox*, 2008]. Therefore, *Steyaert and Knox* [2008] developed a conservative method in order to infer the spatial distribution of potentially water saturated soils for each time slice (for a detailed description of the methodology, see *Steyaert and Knox* [2008, section 2.3.4]). Their objective was to provide a soil moisture boundary condition for land-atmosphere interaction modeling experiments and sensitivity tests on the potential effects of historical changes owing to artificial drainage, including the combined effects owing to changes in biophysical parameters. They derived the fractional area distribution of potential saturated soils (PSS) for each time slice where the PSS is restricted to the peak early growing season (i.e., June) when preceded by normal weather [see *Steyaert and Knox*, 2008, section 3.1.6 and Figure 6]. The 20-km fractional area PSS maps were developed as proxy data to infer the potentially water-saturated soils associated with wetlands vegetation complexes in 1650 and then subsequent changes in PSS that primarily resulted from artificial drainage for agriculture [*Steyaert and Knox*, 2008].

2. Methods

[10] We employed RAMS version 4.4 [*Cotton et al.*, 2003] equipped with the LEAF-2 [*Walko et al.*, 2000] land surface model to examine the effects of historical land cover change in the eastern United States. This modeling system was used to simulate the weather using the 1992, 1920, 1850, and 1650 land cover and biophysical parameter representations. The 1992 run was labeled as the control and its results were compared with observed minimum and maximum temperatures and precipitation. The results of the other three scenarios were compared to those from the control and any differences were assumed owing to the land cover change.

[11] In LEAF-2 the land cover type is represented by a user defined number of patches. Each patch represents the area of the RAMS grid cell covered by a particular land cover type. Four basic components comprise these patches: canopy air, vegetation, temporary surface water (snow), and soil. The canopy air is defined as the air below the vegetation height when vegetation is present. When no vegetation is present the "canopy" air is defined as the air in the viscous sublayer. At each time step, energy and moisture balances are determined for each of these components. The turbulent sensible and latent heat fluxes between vegetation and canopy air are proportional to the total LAI. Similarly, the emission and absorption of longwave radiation and the absorption of shortwave radiation are proportional to the vegetation fraction. The vegetation fraction is defined as the fraction of the ground that is obscured by vegetation when viewed directly from above. Transpiration is influenced by the LAI and the difference between the saturation vapor pressure at the leaf surface and the actual vapor pressure of the canopy air. The transpiration is also

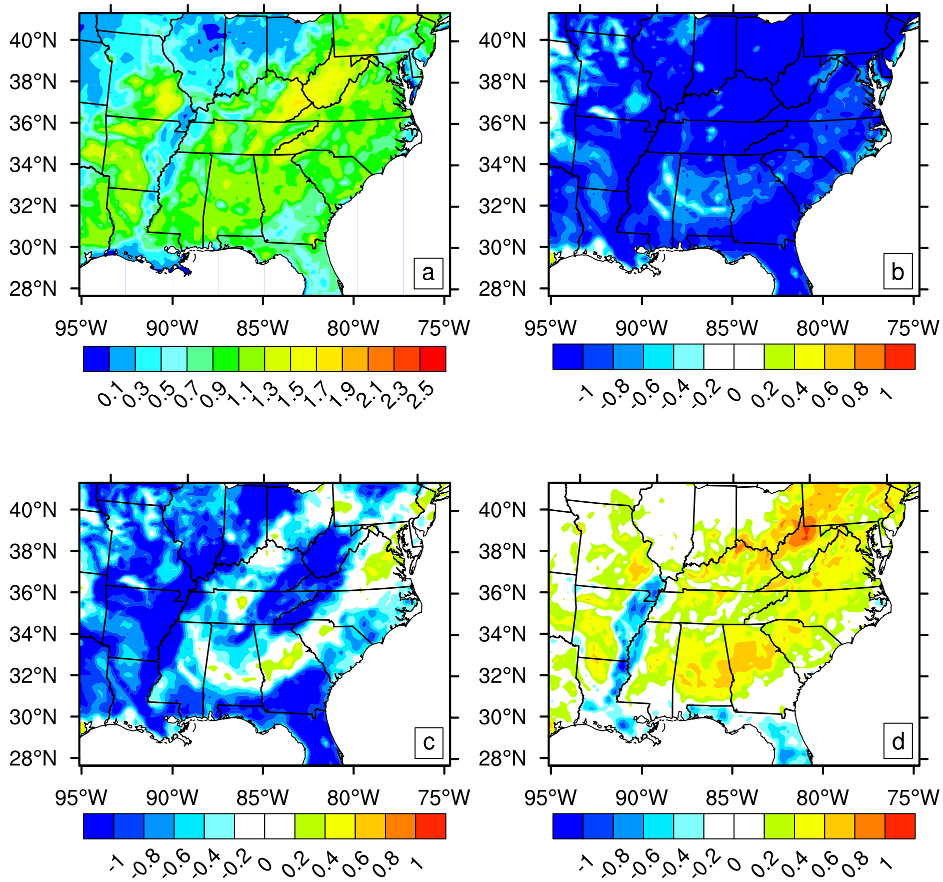

Figure 2. (a) Net surface roughness for 1992. Also shown are differences in surface roughness between 1992 and (b) 1650, (c) 1850, and (d) 1920. Derived from land cover and biophysical parameters [*Steyaert and Knox*, 2008].

related to the stomatal resistance, which is a function of temperature, solar radiation, soil moisture, and vapor pressure deficit at the surface of the leaf. Turbulent sensible and latent heat flux exchanges between the canopy air and lowest model layer are calculated from similarity theory [*Louis*, 1979], and are weighted according to the fractional area of each patch. A detailed listing of the equations used by LEAF-2 can be found in the work of *Pielke* [2002, Appendix D].

2.1. Atmospheric Initialization

[12] Two nested grids were used in this study. The inner grid covered most of the eastern United States and had horizontal grid intervals of 20 km. The outer grid extended past the boundaries of the inner grid by 800 km and had horizontal grid increments of 80 km. Both grids have 42 levels in the vertical extending to an altitude of 24 km. Between the surface and an altitude of 10 km the distance between levels gradually increases from 80 to 1000 m. At altitudes above 10 km the distance between levels is held constant at 1000 m.

[13] The model was initialized at 1200 UTC on 30 May with the NCAR-NCEP reanalysis derived temperature, specific humidity, geopotential height, and winds [*Kalnay et al.*, 1996] and run until 1800 UTC on 30 June. Although the reanalysis is relatively coarse in spatial resolution (grid increments of 210 km), it is adequate for initializing the upper layers of the model. We chose 1996 because, according to the National Climatic Data Center's Time Bias Corrected Divisional Temperature-Precipitation-Drought Index data set [*National Climate Data Center* (*NCDC*), 1994], the June precipitation was close to the 1971–2000 average, and the Palmer Drought Severity Index was very low in the March–June period over much of the eastern United States. An important aspect of this study, of course, is that it provides a conservative estimate of the effect of the landscape change on the weather, since the information that is fed into the models from the lateral boundary conditions is from 1996 weather which developed with the landscape and other climate forcings of this near-current time period. In 1650, 1850 and 1920, however, the atmospheric features would have evolved with different landscapes outside of the modeled domain, in addition to the landscape changes in the interior.

[14] The *Smagorinsky* [1963] and *Mellor and Yamada* [1982] turbulence schemes were used for horizontal and vertical diffusion, respectively. The Smagorinsky scheme calculates the horizontal diffusion coefficients as the product of the local horizontal deformation rate and a length scale proportional to the horizontal grid increment. The Mellor-Yamada scheme uses turbulent kinetic energy simulated from the model's prognostic velocity components to

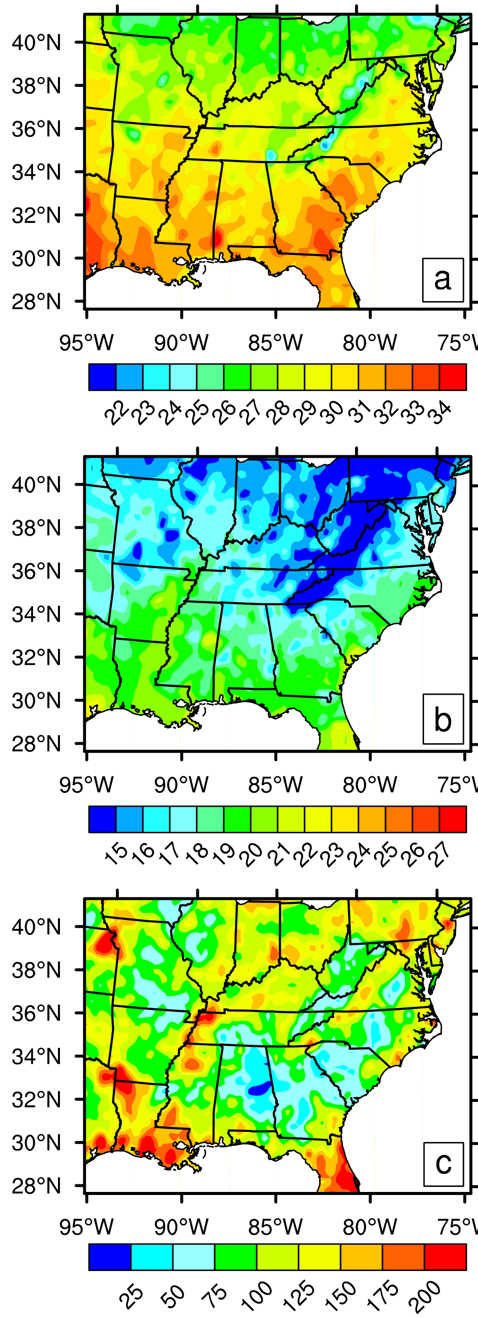

Figure 3. June 1996 (a) USSOD mean maximum temperature (°C), (b) USSOD mean minimum temperature (°C), and (c) CPC gridded USSOD total precipitation (mm). Note that Figure s 3a and 3b are derived from *NCDC* [2006], and Figure 3c is derived from *Higgins et al.* [2000].

calculate the vertical diffusion. Both shortwave and longwave radiation were parameterized using the Chen and Cotton [*Chen and Cotton*, 1983] routine. The Chen and Cotton routine uses the full radiative transfer equation to simulate longwave radiation and a three-band scheme for shortwave radiation. Both schemes account for condensate in the atmosphere, but do not distinguish between cloud water, rain, or ice. The Kuo convective scheme [*Kuo*, 1974; *Molinari*, 1985] was used to simulate precipitation not resolved explicitly by the grid, and a dump bucket scheme [*Cotton et al.*, 1995] was used to simulate large-scale precipitation. The Davies scheme [*Davies*, 1976] was used to nudge the boundaries of the outer grid toward the NCAR-NCEP reanalysis every six hours during the course of the simulation with a timescale of 1800 s. In order to provide a better characterization of the larger meteorological scales [*Rockel et al.*, 2008], very weak internal nudging with a timescale of 1 d was also used on the outer grid.

2.2. Land Cover and Soil Specification

[15] All simulations used the *Steyaert and Knox* [2008] land cover data set which consists of a set of average fractional area land cover maps for each time slice where the fractional land cover values have been averaged over 20-km blocks and provided in geographic coordinates at a 30 arc-second grid increment. The fractional area land cover values sum to 1.0 at each location depending on the number of land cover types from the suite of 36 land cover classes that are needed to characterize the location. Up to nine separate land cover types plus water were allowed in each RAMS grid cell. The number of land cover types actually used in each grid cell varied from region to region and between time slices. The 1650 scenario had the least diverse land cover and generally three to four land cover types would represent the entire area of most grid cells. In contrast, all nine of the land cover types were utilized in some parts of the inner grid domain, especially northern Florida and the coastal sections of Georgia and the Carolinas in the 1992 scenario.

[16] For the simulations at each time slice, the corresponding 20-km PSS data layer (values from 0 to 100%) was used to prescribe the degree of saturated soil according to the PSS value and the land cover type of each patch in the LEAF-2 grid cell. If the PSS was 100% throughout the LEAF-2 grid cell, then soil moisture was held fixed at the saturated level. If the PSS was zero in the LEAF-2 grid cell, then the soil moisture was allowed to vary freely. For intermediate cases of PSS, the fractional area of PSS was allocated to the LEAF-2 patches on the basis of a categorical ranking of PSS and land cover affinity for wetlands (R. G. Knox, personal communication, 2007). The soil in these simulations consisted of 13 levels that extended from the surface to 2.7 m. The layers ranged from 0.03 m thick at the surface to 30 cm thick below 60 cm. The soil textures for all the simulations were assigned on the basis of the analysis of USDA STATSGO data set *U.S. Department of Agriculture* [1994] by *Miller and White* [1998]. The dominant texture from the top 60 cm of the STATSGO data was used for the entire soil profile in the RAM simulations.

[17] The initial soil moisture and temperature were determined from a five month spin up simulation starting at 0000 UTC on 1 January 1996. Soil moisture at the start of the spin up was set equal to field capacity and, because 0000 UTC corresponds to early evening in the region, the soil temperature was set to a constant 2°C above the lowest model level air temperature. The model was then run until 1200 UTC on 30 May and the resulting soil moisture and temperature fields were used to initialize the main run. A separate spin-up was performed for each land cover scenario, i.e., 1650, 1850, 1920, and 1992, because the land cover affects the evolution of the soil temperature and moisture.

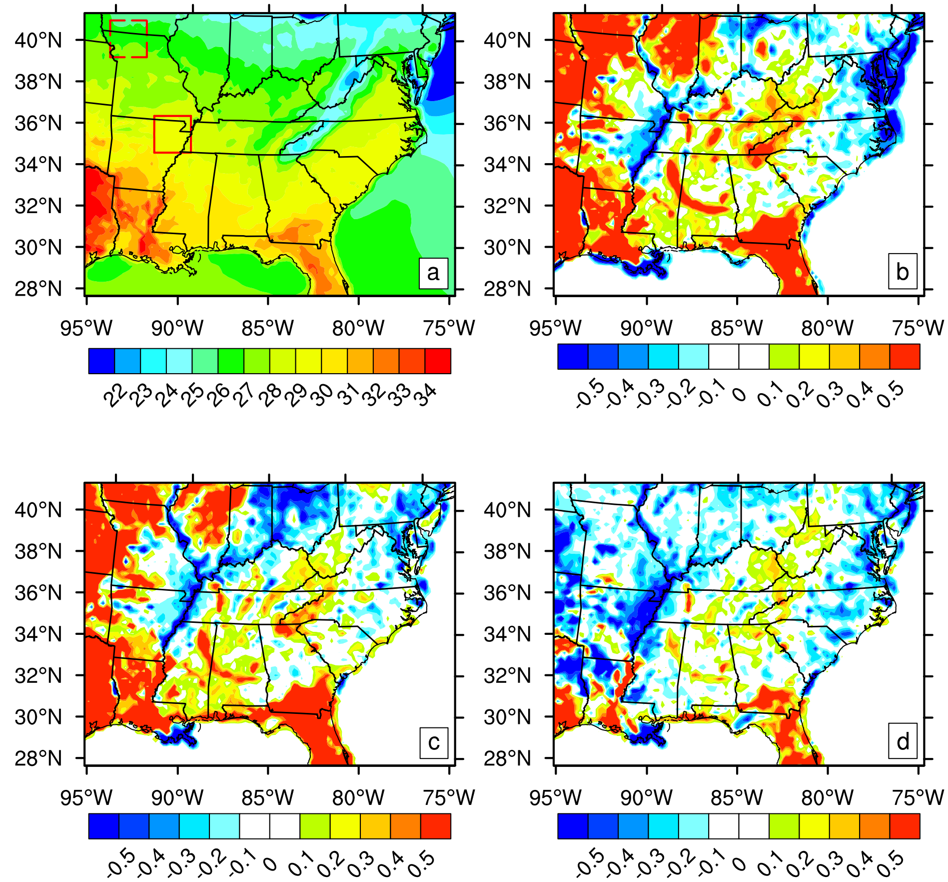

Figure 4. RAMS/LEAF-2 model simulation results: (a) maximum temperature (°C) with 1992 land cover. Also shown are differences in maximum temperature between 1992 and (b) 1650, (c) 1850, and (d) 1920. Note that the dashed and solid boxes in Figure 4a show the subset areas that are used for the diurnal change and biophysical parameter analyses.

Only the upper layers of the soil will have time to adjust to the model in this short spin-up period, but the deeper layers will not have a significant influence on a 1-month simulation.

3. Results

3.1. Maximum Temperatures

[18] Figure 3a shows the observed U.S. Summary of the Day (USSOD) [*NCDC*, 2006] mean maximum high temperatures for June 1996. The high temperatures ranged from 33°C in southeastern Texas to 23°C in the higher elevations of the Appalachian Mountains. Figure 4a shows simulated maximum temperatures for the RAMS/LEAF-2 simulation with 1992 land cover. The root mean square error (RMSE) between the observed and simulated maximum temperatures was 1.6°C. The model does reasonably well at simulating the distribution of temperatures with the warmest temperatures in Texas and the coolest in the Appalachians and along the northern portion of the domain.

[19] Figure 4b shows the difference between the maximum temperatures in the run with the 1992 and 1650 land cover. Warming has occurred in the western portion of the domain while a slight cooling has occurred over northern Indiana, Ohio and along the Mississippi and Ohio rivers. Warming also occurred in portions of the southeastern and Appalachian states. The cooling along the Louisiana coast and along the eastern U.S. coast is mostly a result of an increase in the water fraction of the grid cells in those regions owing to differences in land masks used to develop the data sets. The changes between 1850 and 1992 were similar, while slight cooling occurred along portions of the western domain edge between 1920 and 1992.

[20] In order to explain the changes in temperature we examined the diurnal changes in heat fluxes for two subsets of 100 grid cells, one in a region of daytime warming (region 1 in parts of NW Missouri and SW Iowa) and one in a region of daytime cooling (region 2 in NE Arkansas); see Figure 4a. Table 1 shows the average values of several biophysical parameters in region 1 for each of the land cover scenarios. The land cover in this region was predominantly a mixture of grassland and broadleaf forest in 1650 and 1850. Between 1850 and 1920 most of the region was converted to crops. This change is evident in Table 1 from the decrease in surface roughness, minimum stomatal resistance, LAI, and vegetation fraction as well as the slight increase in albedo. Between 1920 and 1992 a small amount of the crop area was replaced with forest. In addition, to these land cover changes there was small decrease in the area of PSS between 1650 and 1992.

[21] The average diurnal time series of sensible and latent heat flux from the canopy air for the 1992 land cover is shown for region 1 in Figure 5a. Figures 5b and 5c shows

Table 1. Area-Averaged Biophysical Parameters for Region 1

	1650	1850	1920	1992
Surface roughness (m)	1.3	1.2	0.2	0.3
LAI	4.3	4.3	4.0	4.1
Minimum stomatal resistance (s m^{-1})	143	140	115	116
Land cover albedo	0.16	0.16	0.18	0.18
Vegetation fraction	0.86	0.86	0.84	0.84
Root depth (m)	2.3	2.3	1.0	1.0

the differences between 1992 and each of the other land cover scenarios. Figure 5b illustrates that with the 1992 land cover, the sensible heat flux is initially larger in the morning hours but then becomes smaller during the middle of the day. The latent heat flux shows a similar pattern, although the differences have greater magnitudes. These differences are primarily explained by the changes in surface roughness length and minimum stomatal resistance. The initial higher sensible heat fluxes are due to decreased morning cloud cover in the 1992 scenario. The lower minimum stomatal resistance in 1992 causes the transpiration to be greater than in 1650. As the morning progresses, the difference in solar radiation between 1992 and 1650 decreases. In addition, as the wind speed increases during the middle of the day, the effect of the surface roughness on the sensible and latent heat flux increases. After midday the sensible heat flux for the 1992 land cover begins to approach, and eventually becomes greater, than that with the 1650 land cover. This occurs because the evaporation from the soil in 1992 is less than in 1650, owing to the reduced surface roughness and PSS. The reduced soil evaporation allows the soil and canopy air to warm to greater temperatures and thus leads to the enhanced canopy sensible heat flux and a reduction in latent heat flux from the canopy air. Over the course of the day the cumulative sensible heat flux for 1992 is 849 W m^{-2} compared with 815 W m^{-2} for 1650. This leads to the warmer daytime high temperatures shown in Figure 4b. The differences between 1850 and 1992 are similar to those for 1650 since there was little change in land cover between 1650 and 1850 in that region. The maximum temperatures declined somewhat between 1920 and 1992. This is

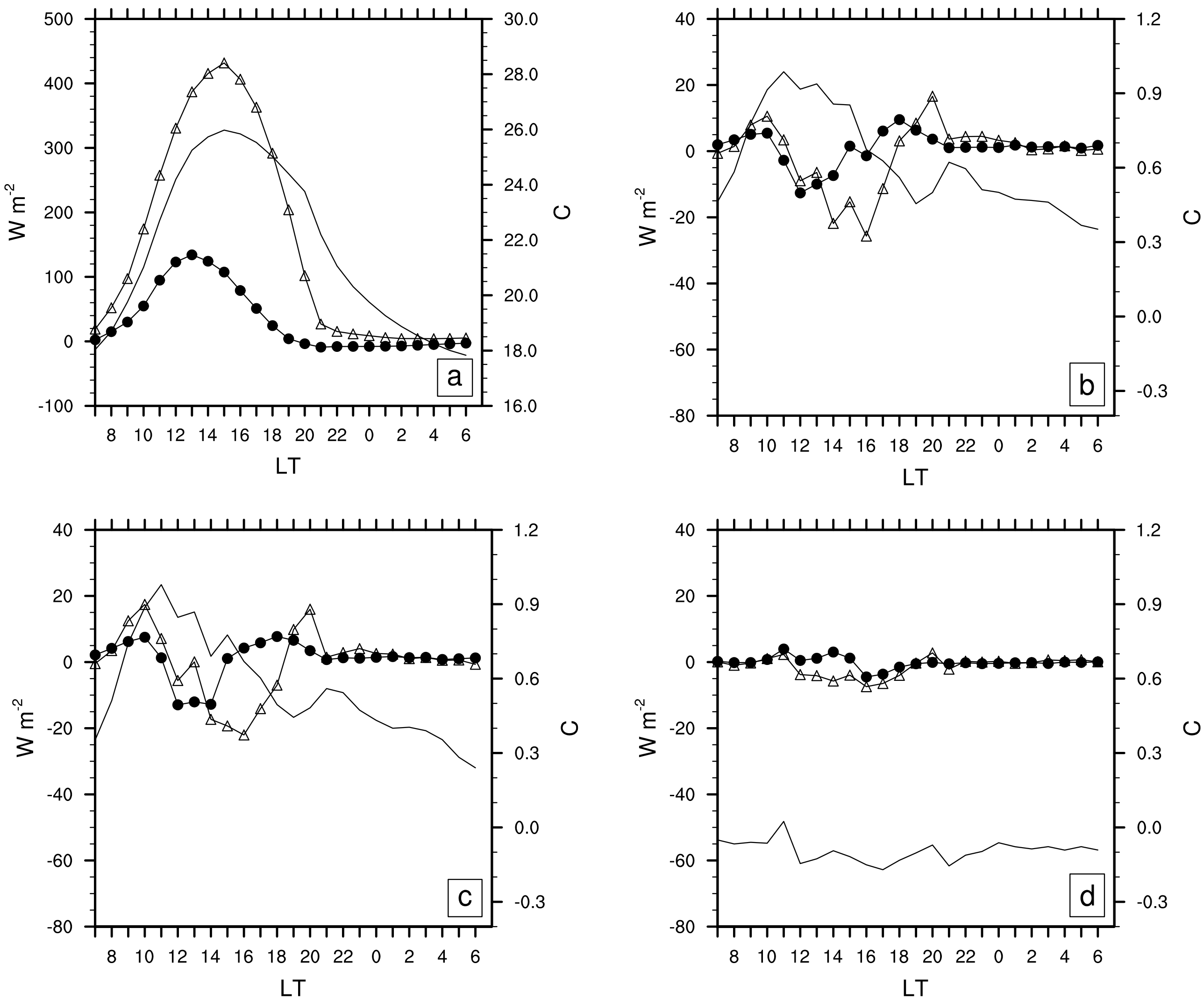

Figure 5. (a) Mean sensible heat flux (solid circles), latent heat flux (triangles), and air temperature (unmarked) for region 1 with 1992 land cover. Also shown are differences in sensible and latent heat fluxes and air temperature between 1992 and (b) 1650, (c) 1850, and (d) 1920.

Table 2. Area-Averaged Biophysical Parameters for Region 2

	1650	1850	1920	1992
Surface roughness (m)	2.3	2.2	0.9	0.6
LAI	5.0	5.0	4.4	4.3
Minimum stomatal resistance (s m^{-1})	487	452	208	161
Land cover albedo	0.15	0.15	0.17	0.17
Vegetation fraction	0.90	0.89	0.85	0.86
Root depth (m)	2.0	2.0	1.4	1.3

expected since the area of crops shrunk slightly during this time period.

[22] Table 2 shows the average values of several biophysical parameters in region 2 for each of the land cover scenarios. The land cover in this area was mostly forest in 1650 and 1850. By 1920 about a third of the area had been converted to crops and much of the remaining forest consisted of trees with lower heights than in 1650 and 1850. By 1992 about 60% of the area had been converted to crops. Table 2 shows that this change has led to large decreases in surface roughness and minimum stomatal resistance. Vegetation fraction and LAI also declined while the albedo increased slightly. Finally, there was also a significant drop in PSS in region 2 between 1650 and 1992.

[23] Figure 6a shows the diurnal cycle of the sensible and latent heat flux from the canopy air for Region 2 with the 1992 land cover. From Figure 6b it can be seen that the morning sensible heat flux is slightly greater in 1992 than 1650. This is again due to slightly less cloud cover and more solar radiation. The latent heat flux is increased in 1992 owing to the much greater transpiration. The drop in minimum stomatal resistance between 1650 and 1992 in this region is more pronounced than in region 1, and the increased transpiration more than offsets the drop in evaporation from the soil owing to the decreased surface roughness and PSS. Similar results are found for the 1850 and 1920 cases. The differences are not as large in 1920, however, since a portion of the change over to crops has already occurred by then. One should note that in dry years the relative coverage of PSS may have a more significant impact if non-PSS areas dry out enough to make transpiration negligible. This pattern of reduced daytime sensible heat flux leads to the reduction of high temperatures shown in Figures 4b–4d. The other areas of warming and cooling

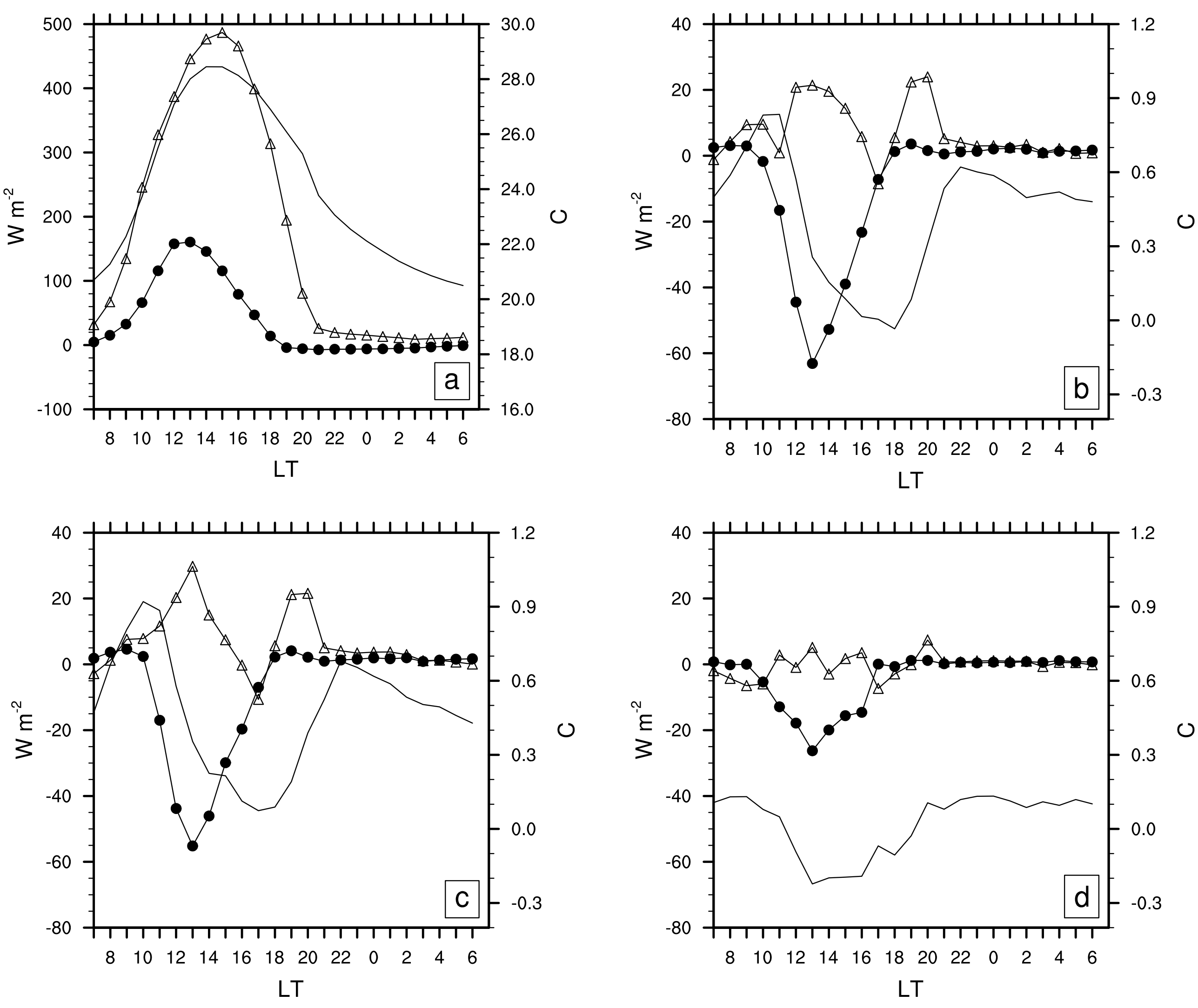

Figure 6. Same as Figure 5, except for region 2.

Table 3. Mean and Standard Deviation of Temperature Difference Between 1992 and Earlier Land Cover Scenarios

	1992–1650	1992–1850	1992–1920
Maximum temperature (°C)	0.3 ± 0.6	0.3 ± 0.6	0 ± 0.3
Minimum temperature (°C)	0.4 ± 0.2	0.2 ± 0.2	−0.1 ± 0.2

of the maximum temperatures across the domain are explained by similar changes in surface roughness and stomatal resistance. Table 3 shows the average difference between 1992 and the other years in maximum temperatures at all land points in the domain. When compared with 1650 and 1850 the 1992 high temperatures are slightly warmer, while they remained about the same when compared with 1920.

3.2. Minimum Temperatures

[24] Figure 3b shows the observed USSOD mean minimum temperatures [*NCDC*, 2006] for June 1996. The low temperatures ranged from 21°C along the Gulf and southeast coasts to less than 15°C in the higher elevations of the Appalachian Mountains. Figure 7a shows simulated minimum temperatures for the run with 1992 land cover. The RMSE between the observed and simulated minimum temperatures was 1.4°C. The model does reasonably well at simulating the distribution of temperatures with the warmest temperatures along the coasts and the coolest in the Appalachians and along the northern portion of the domain. The model also simulates the tongue of warmer minimum temperatures that extends up the Mississippi valley.

[25] Figure 7b shows the difference between the minimum temperatures in the run with the 1992 and 1650 land cover. Slight warming has occurred over almost all the domain. The warming along the Louisiana coast and along the eastern U.S. coast is, like the cooling in maximum temperatures, mostly a result of an increase in the water fraction of the grid cells in those regions owing to differences in land masks used to develop the data sets.

[26] The explanation for the warming at night is simpler than that for the daytime temperatures. The temperature at night takes longer to fall in the regions where daytime temperatures warmed owing to the extra accumulation of heat during the day. In the regions that experienced daytime cooling owing to increased transpiration the nocturnal temperatures warmed because of the extra water vapor reducing outgoing longwave radiation. The minimum temperatures are also warmer in most areas when compared to 1850. A few areas in Virginia and Pennsylvania are slightly cooler in 1992 than 1850 at night. This is where daytime temperatures also cooled. Finally, Figure 7d shows that the low temperatures are cooler in 1992 than 1920 over much of the domain. This is due to the slight daytime cooling that occurred during this time period. When compared with 1650 and 1850 the domain-average minimum temperatures

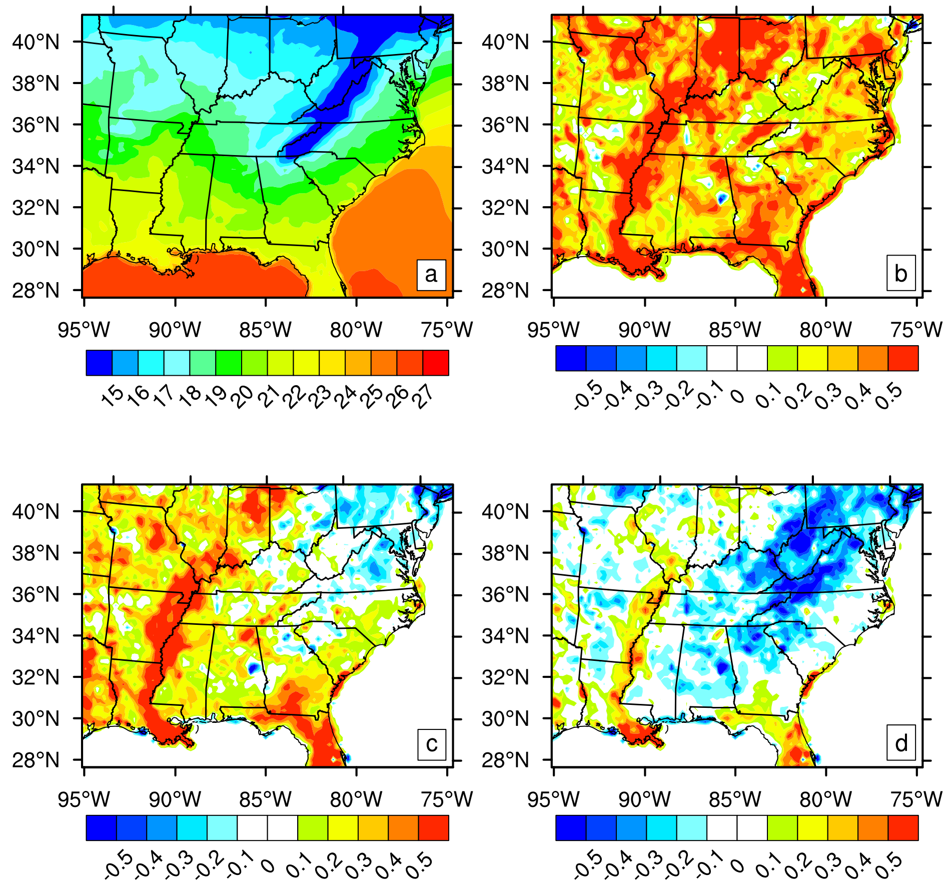

Figure 7. RAMS/LEAF-2 model simulation results: (a) minimum temperature (°C) with 1992 land cover. Also shown are differences in minimum temperature between 1992 and (b) 1650, (c) 1850, and (d) 1920.

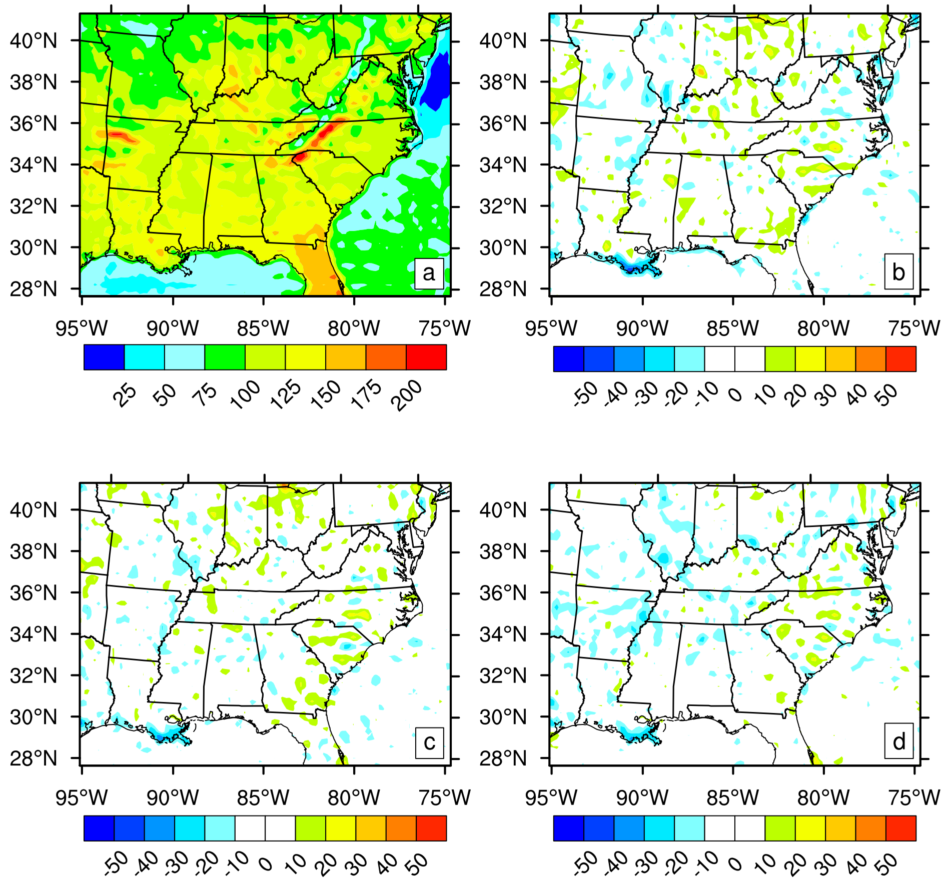

Figure 8. RAMS/LEAF-2 model simulation results: (a) precipitation (mm) with 1992 land cover. Also shown are differences in precipitation between 1992 and (b) 1650, (c) 1850, and (d) 1920.

over land in 1992 are slightly warmer, while they are slightly cooler when compared with 1920; see Table 3.

3.3. Precipitation

[27] Figure 3c shows the NOAA Climate Prediction Center (CPC) 0.25-degree gridded Daily U.S. Unified Total Precipitation totals [*Higgins et al.*, 2000] for June 1996. This data set was derived from a combination of the National Climate Data Center cooperative observer stations, the CPC precipitation stations, and the daily sums from the hourly precipitation data set. Figure 3c shows the data re-gridded to the 20 km RAMS grid used in this study.

[28] The greatest precipitation, in excess of 200 mm, occurred over Louisiana, Florida, and northeastern Kansas. The driest area, with less than 75 mm, was over portions of Alabama and Georgia. Figure 8a shows the simulated precipitation for the 1992 land cover case. The model was often within 25 mm in the northern part of the domain, but overestimates significantly in the observed dry region in Alabama and Georgia. Also, the model underestimates the observed maxima in Louisiana and northeastern Kansas. The RMSE between the observed and simulated precipitation was 41 mm. Figures 8b–8d show that the land cover changes between the different years had little effect on precipitation.

3.4. Sensitivity to Soil Moisture

[29] Three additional runs were made in order to determine the sensitivity to soil moisture. In the first two runs the soil wetness was reduced by 0.3 across the entire domain for 1992 and 1650; see Figure 9. The change in 2-m air temperature between 1650 and 1992 was very sensitive to this change in soil moisture, with both daytime and nighttime values becoming warmer. When the 1992 temperatures are compared with those for 1650 there were almost no areas of cooling. Even the area along the Mississippi River, which cooled between 1650 and 1992 when the spun-up soil moisture is used, exhibited warming when the soil moisture is reduced. Averaged over all land points the maximum temperatures warmed by 2.1 ± 1.3°C and the minimum temperatures warmed by 0.5 ± 0.2°C between 1650 and 1992 when the reduced soil moisture is used for both of those years. However, when compared to observed values, the temperatures from the 1992 run with reduced soil moisture were often much too warm and the results from the spun-up soil moisture run are more realistic. It should be noted that the soil moisture evolution in these simulations is strongly influenced by the accuracy of the precipitation produced by the Kuo convective scheme. The results from the control run show that the scheme significantly overestimates the precipitation in the southeastern United States, which likely leads to an overestimate of the soil moisture.

[30] The third run used the spun-up soil moisture but with the areas of PSS removed for 1650. When compared against the 1650 run containing PSS, this had little effect on temperature or precipitation, apparently because of the

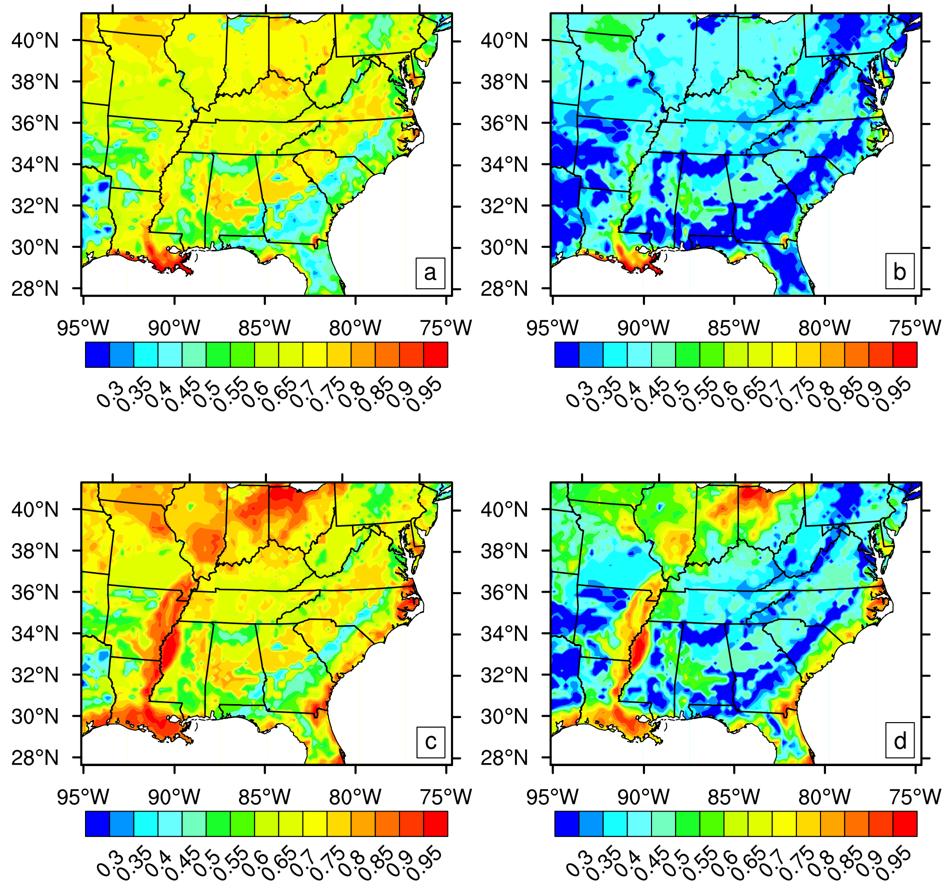

Figure 9. RAMS/LEAF-2 model simulation results: initial soil wetness (percent saturation) for (a) 1992, (b) 1992 with 0.3 reduction in wetness, (c) 1650, and (d) 1650 with 0.3 reduction in wetness.

relatively small area of the domain affected by this change. The PSS may play a more important role during dry periods if those areas remain moist while non-PSS areas dry out. For example, recent work by *Koster et al.* [2004] suggests that transition zones between low and high soil moisture can serve as focal points for precipitation. More simulations are needed to determine the importance of PSS to near-surface climate.

4. Conclusions

[31] In this study RAMS with LEAF-2 was used to investigate the influence of land cover change on summer minimum and maximum temperatures, and on precipitation. We simulated the weather over the eastern United States for the month of June 1996, a month of near average temperature and precipitation, using the *Steyaert and Knox* [2008] land cover for 1992, 1920, 1850 and 1650. These time slices captured the replacement of the eastern forests with crops between 1650 and 1920, and the subsequent partial regrowth of the forest since 1920. The modeled June weather appeared to be most sensitive to changes in surface roughness and minimum stomatal resistance. Reductions in surface roughness acted to reduce the magnitudes of both sensible and latent heat fluxes, however, in some cases the reduction in evaporative cooling caused by the decreased latent heat flux actually allowed the sensible heat fluxes to become larger when the surface roughness was decreased. Reductions in minimum stomatal resistance acted to increase latent heat flux and thus decrease sensible heat flux. In most areas of the domain, replacement of forest with crops led to decreases in surface roughness and minimum stomatal resistance and a warming of both daytime and nighttime temperatures. During the day the reduced surface roughness decreased the latent heat flux and thus evaporative cooling of the surface. In addition, morning cloud cover decreased slightly over most of the regions with reduced roughness leading to a slight increase in solar radiation. This led to increased sensible heat flux and warmer temperatures. The warmer temperatures during the night appear to be due to the extra heat gained during the day and increased atmospheric water vapor.

[32] In a few areas along the Mississippi River and in northern Indiana and Ohio the forest to crop transition led to cooling of the daytime temperatures. This occurred because the drops in stomatal resistance were large enough to boost transpiration to sufficient levels to compensate for the loss of evaporative cooling owing to the reduction in surface roughness. Also, a significant drop in the LAI in these areas reduced the sensible heat flux from the vegetation, which further enhanced the daytime cooling.

[33] We found that when averaged over all the land points in the domain both maximum and minimum temperatures warmed slightly when compared with the 1650 and 1850 scenarios, while maximum temperatures remained about the same and minimum temperatures cooled slightly when compared with 1920. There was little change in precipitation. These small responses of near surface air temperature

to the changes in land cover are comparable to those calculated by *Copeland et al.* [1996] and *Baidya Roy et al.* [2003], as well as from a modeling study of land cover change effects in temperate and subtropical South America by *Beltrán* [2005], and for Australia by *Narisma and Pitman* [2003]. Other studies, summarized in reviews including *National Research Council* [2005], *Kabat et al.* [2004] and *Cotton and Pielke* [2007], find similar magnitude effects of landscape change on temperature.

[34] A separate set of runs where the initial soil moisture was reduced suggests that the land cover change effects on temperature could be magnified in dry years. When the temperatures from the 1992 and 1650 runs with reduced soil moisture were compared, both daytime and nighttime values increased across all of the domain and, especially during the day, by a significantly larger amount. This was likely due to reduced moisture available for evaporative cooling either by transpiration or soil evaporation.

[35] The results from this study suggest that under average to slightly above average soil moisture conditions the changes in land cover that have occurred across the eastern United States since 1650 have had an influence on minimum and maximum temperatures of generally 0.5°C or less. The effects appear to be stronger under drier soil conditions with daytime temperatures as much as 2°C warmer during the day. Since 1920 the temperatures have cooled slightly owing to regrowth of the forest over many areas. It should be noted, however, that we have only simulated one month from a single year. The magnitude and direction of these changes could vary greatly between seasons and in years with significant departures from average precipitation and temperature. In addition, model simulations where individual biophysical parameters thought responsible for the changes described above are held fixed could be run, although this is computer and time intensive. Thus these experiments therefore, are deferred to later study. Such a perturbation type of experimental design would examine other mechanisms for the near surface temperature behavior. Nonetheless, we show clearly that landscape changes do have a significant effect on regional and local climate, and must be considered in any assessment of the role of humans in climate variability and change.

[36] Another caveat that should be mentioned is the potential sensitivity of these findings to the land surface parameterizations used. For example, *Oleson et al.* [2004] showed how the choice of land cover scheme influenced the amount of simulated summer cooling owing to land cover change in the north central United States. They found in particular that the amount of cooling was sensitive to differences in the treatment of transpiration and soil evaporation processes between two land surface models. Owing to the small magnitudes of the changes shown in our study it is possible that use of a different land surface model could reverse the direction of the change in some cases. Despite these shortcomings our study further supports the conclusion of *Hale et al.* [2006], that when examining multidecadal trends in the surface temperature record, for regions of significant land cover change, one should consider the effects of those changes in land cover during the period. These changes are superimposed on those brought about by other human and natural climate forcings.

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